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CORRECTIONS AND EMENDATIONS

Contributors to Volume 54 have been invited to send corrections and emendations to be made in their papers, and the volume has been scanned with some care. Corrections and insertions are as follows:

Pages 99 and 113.	Pictures for Figure 1 of Plate 3 and Figure 1 of Plate 9 are exchanged.
Page 321, No. 68.	*For* Narragansett, *read* Cumberland.
" 571, Figure 2.	In profile of Hartford Section, omitted gravity anomaly for Station 632 is — 7.
" 583, Table 2.	Last column, *for* 2.53, *read* 2.55.

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JANUARY 1, 1943

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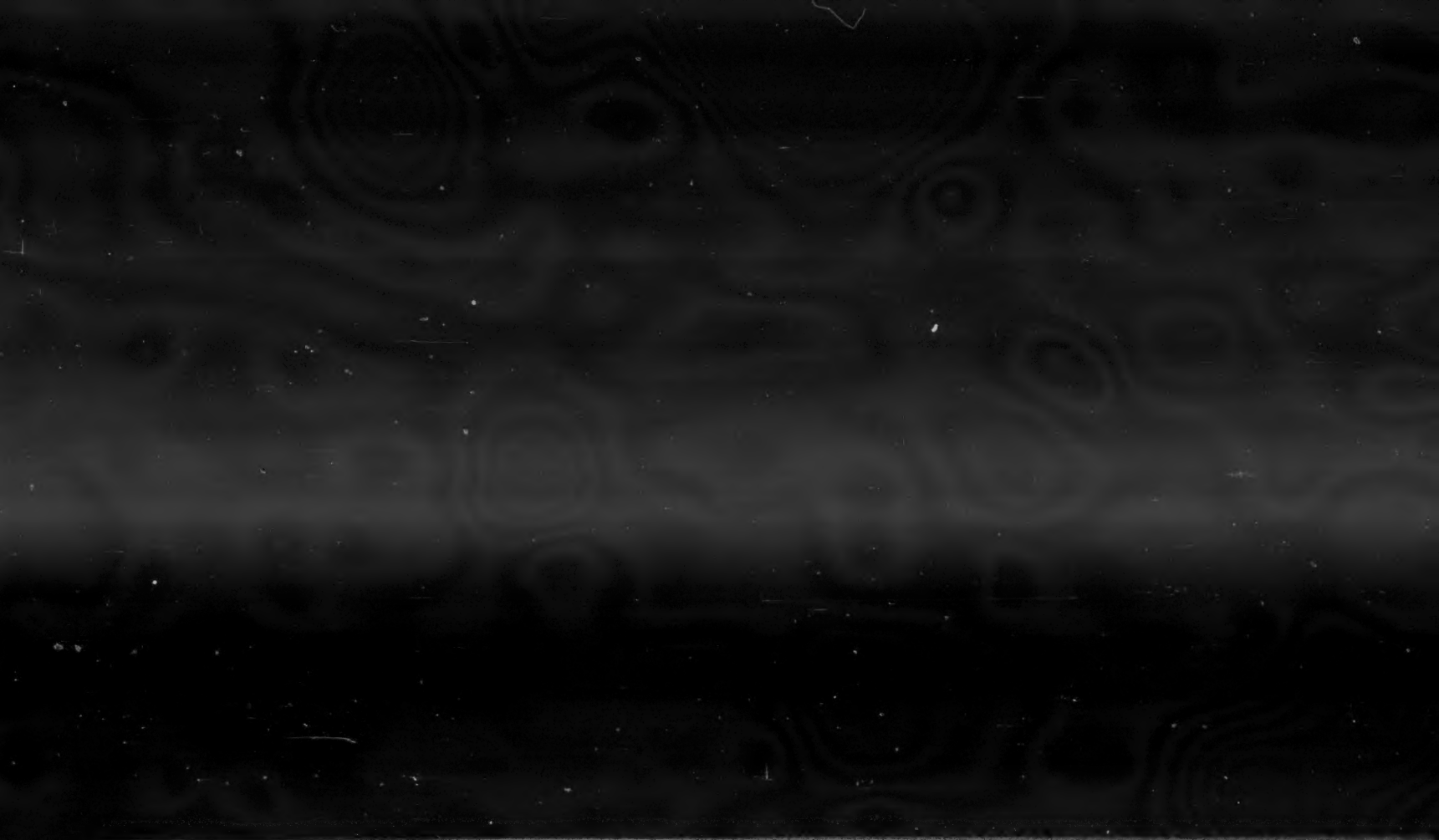

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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STRATIGRAPHY AND STRUCTURE AT THREE FORKS, MONTANA

BY GEORGE W. BERRY

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ABSTRACT

The northwest part of the Three Forks sheet, mapped by Peale (1896), has been remapped. The Pebbly limestones (Cambrian) of Peale's section could not be identified. The contact of the Dry Creek shales with the overlying Jefferson limestone represents a hiatus involving the Ordovician, Silurian, and Lower Devonian. The Jefferson is correlated with the Devils Gate of Nevada; upper Middle Devonian age is indicated in the lower beds by *Spirifer* cf. *S. engelmanni (Spirifer argentarius* zone, Devils Gate), and the base of the Upper Devonian is marked by the occurrence of *Phillipsastraea,* 175 feet above the base of the formation. Sixty feet of yellow sandstone overlying the *Cyrtospirifer* zone of the Three Forks shale (Upper Devonian), and previously included in that formation, contains a Lower Carboniferous *Syringothyris* fauna (Kinderhook) and is lithologically a recognizable unit. The name Sappington sandstone is here proposed. The Belt is not present in the southern part of the area but is represented by 4000 feet of conglomerate, arkosic sandstone, and micaceous shale in the northern part. A relatively steep shore line is indicated, and the section probably is the near-shore facies of one or more of the formations distinguishable farther north. The Belt shore line is marked by an east-west zone of Laramide high-angle thrusting, including the Jefferson Canyon fault, with an apparent throw of approximately 12,000 feet. The faulting was probably controlled by the declivity in the basement surface, which in turn was localized by a zone of weakness within the basement. North of the Jefferson Canyon fault, the structures conform to those of the Northern Rockies in their orientation and plainsward thrusting. South of it, the trends are west and northwest, and thrusting and overturning are to the southwest. This conforms to the general structural pattern of southern Montana and is not explainable by local resolution of east-northeastward Laramide stresses.

INTRODUCTION

The Three Forks sheet, Montana (Peale, 1896), represents an area of 1 square degree. The northwestern one eighth, including most of the region between 111° 30′ and 112° and between 45° 42′ and 46° (Pl. 1; Fig. 1), considered in this discussion, lies in the east-central part of the Northern Rocky Mountain province (Fenneman, 1930, map), and comprises approximately 450 square miles in southeastern Jefferson, southern Broadwater, northeastern Madison, and northwestern Gallatin counties.

The Three Forks area is named for the junction of the Jefferson, Madison, and Gallatin rivers forming the head of the Missouri River. The Madison flows north along the eastern boundary of the area mapped. The Jefferson flows east and northeast through the central part of the region.

From an elevation of about 4000 feet, at the head of the Missouri River (Fig. 1), hills rise to about 5300 feet in the eastern half of the area. However, in the more rugged western portion, the average elevation of the

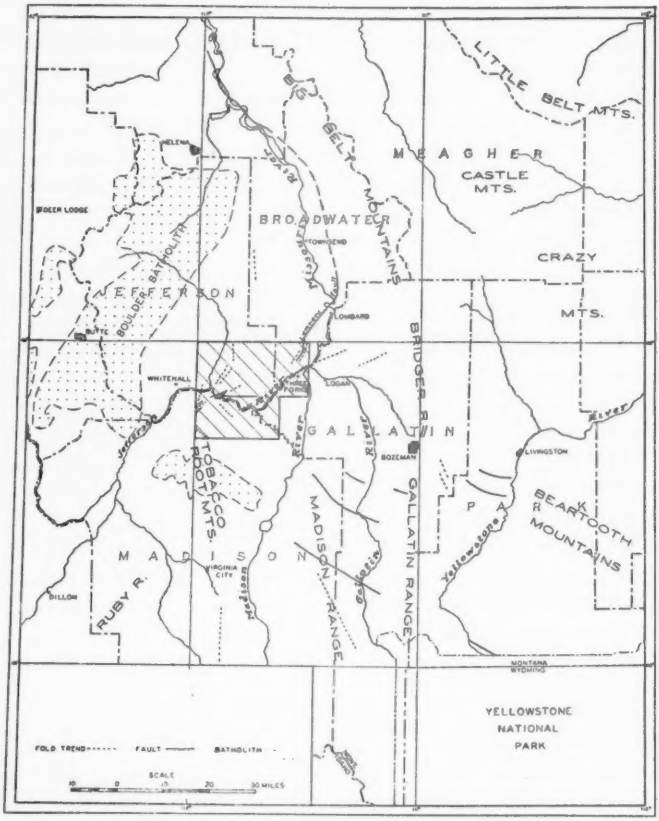

FIGURE 1.—*Structural trends in southwestern Montana*

summits is over 6000 feet, and, in the southwest, peaks in the northern Tobacco Root Mountains attain altitudes of 7800 feet.

The writer spent 20 weeks in the summers of 1939 and 1940 in the field. The base map was constructed from the U. S. Geological Survey Three Forks topographic sheet, County and Forest Service maps; and from aerial photographs (covering the area of the map east of the center of

townships 1, 2, and 3 N., R. 1 W., and east of the center of T. 1 S., R. 2 W.). Where possible, section corners were used in plotting the points in the field. Stratigraphic sections were measured by pace and compass methods.

PREVIOUS WORK

The earliest geologic examination of the area at the head of the Missouri River was made in 1860 by Hayden (1869, p. 90), with the Raynolds expedition. This work was necessarily very general, but accompanying Hayden's report is a geologic map for the entire State east of the Continental Divide. Hayden revisited the Three Forks region (1872, p. 140-142; 1873, p. 82), when he made further general geologic observations. Collections of "Silurian", Carboniferous, and Cretaceous fossils were identified and described by Meek (1873).

In 1883, Davis (1886), Lindgren (1886), Eldridge (1886), and others, in connection with the Northern Transcontinental Survey, under Pumpelly, examined a large area east and north of Three Forks. Their map (Pls. 50-54) represents the topography of the area considered in the present paper and shows the general geology of the region east of Boulder River and north of the latitude of Willow Creek. However, the texts of these reports contain no references to the area west of Three Forks.

In 1884, Peale (1893) began work in the vicinity of Three Forks and north of the Gallatin River. He named and described the formations and constructed a geologic map[1] and cross sections. Merrill (1893; 1895) described a number of the rock specimens collected by Peale. This work resulted in completion of geologic mapping for the Three Forks folio (Peale, 1896). All Peale's maps for this region are reconnaissance.

The area west of Jefferson Island is represented on Weed's geologic map on the Butte District (1912, Pl. 1). The map, however, is no more detailed than Peale's earlier one. In the region west of the Missouri River and north of Three Forks, Haynes (1916a) mapped the geology, measured several Paleozoic sections, and named and discussed the Lombard overthrust and related structures. Condit (1918, p. 115, Pls. 9, 10; Fig. 14; Condit, *et al.*, 1928, p. 179-181, Pl. 12) described sections and mapped the distribution of the Phosphoria formation in the Three Forks area. The general geology of the area is also represented on a map accompanying the report of Pardee and Schrader on the Greater Helena mining region (1933, Pl. 1), and on Billingsley's map of the Boulder batholith and vicinity (1915, Pl. 1). The gneisses and schists of the northern Tobacco Root Mountains have been described and mapped by Tansley and Schafer (1933, p. 8-9, Pl. 1; Tansley, 1936, p. 23-27).

[1] Bounded by 111° 5′ and 111° 40′ and by 45° 45′ and 46°.

The Devonian section of the Three Forks region has been studied. Raymond (1907; 1909; 1910) and Haynes (1916b) have done detailed stratigraphic and paleontologic work on the Three Forks formation. Kindle (1908b, p. 8-9) has examined the Jefferson formation and its fauna northeast of Logan.

Scott (1935) measured several sections of the Amsden and Quadrant formations north and west of Three Forks and discussed the stratigraphic relations and age of these beds.

The stratigraphic section in the vicinity of Three Forks was one of the first described in Montana, and the area contains the type localities of a number of formations. However, Peale's work was largely reconnaissance, so that more exact and complete description of the stratigraphy and paleontology was desirable. Portions of this important section have since been described in detail. The purpose of this study is to re-examine the geology both here and in adjacent regions and to contribute new data regarding the structure and stratigraphy.

ACKNOWLEDGMENTS

To Dr. C. M. Nevin the writer is grateful for assistance in the field in connection with the structural problems, and for assistance in the preparation of this paper. Acknowledgment is also made to Dr. C. W. Merriam, who suggested the area for study, examined the Devonian and lower part of the Carboniferous section in the field, and greatly aided in the preparation of the section on stratigraphy.

STRATIGRAPHY

PRE-CAMBRIAN

Pony series.—The oldest rocks exposed in the Three Forks area are the pre-Cambrian gneisses and schists of the Pony series. (Tansley and Schafer, 1933, p. 8-9; Tansley, 1936, p. 23-27.) Although lithology varies widely and the structure is complicated, in this discussion the series is treated simply as basement complex.

There is a major area of basement in the southern part of the region (Pl. 1), although it is covered in part by Tertiary and recent sediments in Willow Creek valley. Another much smaller exposure occurs on the upthrow side of a major fault about 4 miles northwest of Sappington. These are the northernmost exposures of the core of the Tobacco Root Mountains.

Belt series.—The oldest sedimentary rocks are assigned to the Belt series. A measured section on the north side of Jefferson valley, 4 miles

northwest of Sappington, shows approximately 4000 feet of dark, greenish-gray, coarse, arkosic sandstones. No faulting or repetition was apparent in this section. Locally, the sandstones are conglomeratic, containing

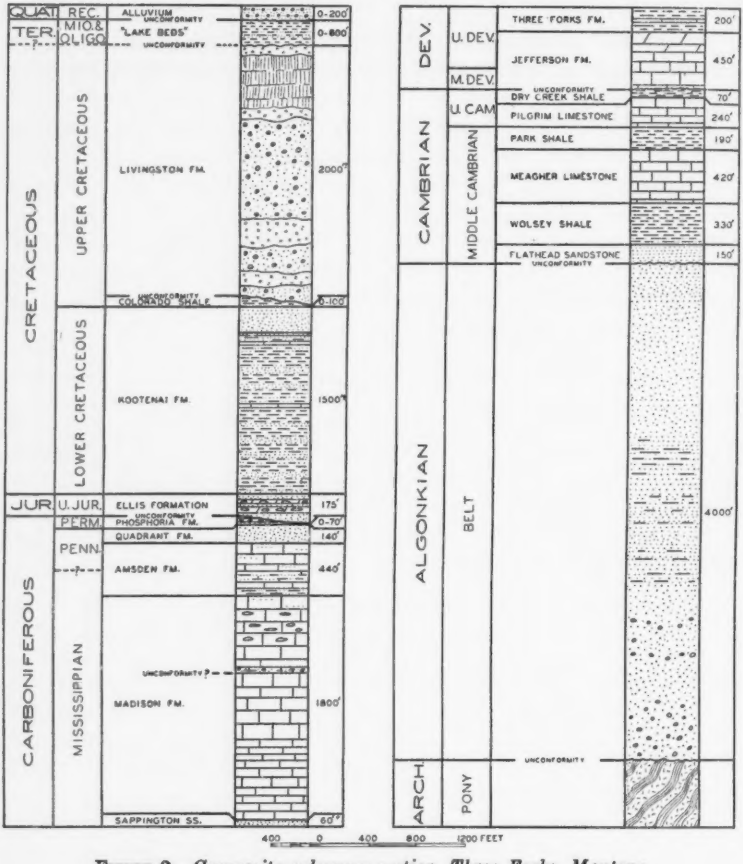

FIGURE 2.—*Composite columnar section, Three Forks, Montana*

pebbles of gneiss and white vein quartz. In the middle portion of the section micaceous shales and sandstones predominate.

A profound unconformity separates the Belt from basement rocks. The contact with the overlying Flathead sandstone (Middle Cambrian) is marked by a sharp lithologic change, but no observed angular discordance.

At a number of localities to the north and northeast, however, angularity has been reported (Deiss, 1935, p. 119-120). Deiss (1935, p. 124) stated that the magnitude of the unconformity separating the Belt and Flathead in western Montana is proof of pre-Cambrian age. Hinds (1935, p. 44-45) has suggested, however, that the unconformity is not nearly so great as Deiss postulated.

Although Belt rocks are not present upon the basement complex 2 miles south of Sappington, 3 miles north of the village they are about 4000 feet thick. Because of this, and similar relations in the area to the east, Peale (1893, p. 14) suggested that a pre-Cambrian shore line extended from the Bridger Range westward, crossing the Madison and Jefferson Rivers 10 or 12 miles south of their junction. However, Deiss (1935, p. 111) considered that such a variation in thickness in less than 5 miles plus the fact that the Flathead sandstone lies upon both the Beltian and "Archean" rocks within the same distance

> ". . . strongly suggest that at the end of Beltian time, deposits of that period extended much farther south than their present boundary would indicate, and also that they and much of the old Archean island were removed by erosion before the invasion of the Middle Cambrian (Flathead) sea."

Field evidence appears to favor Peale's conclusions as to the position of this shore line. No Belt rocks are found to the south, in the Tobacco Root, Madison, and Gallatin ranges. Immediately north of the shore line the sediments are largely coarse clastics, in which quartz and unweathered feldspar grains are but slightly rounded. At Jefferson Canyon, conglomerates containing gneissic boulders 18 inches across prove that the shore line was close and relatively steep during the deposition of these beds. However, the coarse conglomerates are probably restricted to the lower part of the section.

The uniform thickness of the Flathead sandstone throughout the area indicates a nearly level surface upon the Belt and older rocks. Deiss (1935, p. 111) designated this surface as a "peneplain" resulting in a "tremendous amount of erosion" prior to the invasion of the Flathead sea. However, the alternative explanation that the Belt rocks were not deposited far south of their present limit, and that the basement complex was considerably reduced by erosion which supplied the Belt sediments, seems to fit the field evidence somewhat better. Thus, the amount of pre-Flathead erosion would not necessarily have been very great.

North of Three Forks, Haynes (1916a, p. 274-275) recognized the Spokane and Empire formations of the Belt series. Deiss (1935, Pl. 8) figured the Belt rocks of the Three Forks area as the basal portion of the Algonkian section—Neihart, Pritchard, Ravalli, and Newland formations, but in Jefferson valley, lithologic differences are not pronounced enough

to use as a basis for subdivision of the series. Detailed correlation of the Belt section in this region with that of the Little Belt Mountains is inadvisable, since, at present, it would necessarily be based largely on lithology. Moreover, the Belt sediments exposed in Jefferson valley probably represent the near-shore facies of one or more of the formations more easily distinguishable to the north and northwest.

CAMBRIAN

Middle and Upper Cambrian are represented by approximately 1400 feet of strata, mostly limestone and shale. Peale (1893, p. 20-25; 1896, p. 2) described seven divisions of the Cambrian system. Two lower members, the Flathead quartzite and the Flathead shales, were included in the Flathead formation. Five upper members—Trilobite limestones, Obolella shales, Mottled limestones, Dry Creek shales, and Pebbly limestones—constituted the Gallatin formation. To the seven divisions of the Cambrian described by Peale, Haynes (1916a, p. 276-278) applied formation names used by Weed and Pirsson (1900, p. 284-287) in the Little Belt Mountains—Flathead quartzite, Wolsey shale, Meagher limestone, Park shale, Pilgrim limestone, Dry Creek shale, Yogo limestone—but he retained the broader divisions, Flathead and Gallatin formations, for purposes of mapping.

Deiss (1936, p. 1311-1317) carefully measured the Cambrian section of Nixon Gulch, about 10 miles east of Three Forks and 2 miles north of the Gallatin River. He described six formations: Flathead sandstone, Wolsey shale, Meagher limestone, Park shale, Pilgrim limestone, and Dry Creek shale. These correspond to the lower six members of the Flathead and Gallatin formations of Peale and Haynes. Deiss (1936, p. 1316-1317) stated that the upper member of Peale's section—Pebbly limestones (Yogo limestone of Haynes and others)—does not exist. Deiss (1936, p. 1341; 1938, Fig. 7; 1939, Fig. 6) and Bell (1941, p. 198) tentatively assigned the Flathead, Wolsey, Meagher, and Park formations to Middle Cambrian, and the Pilgrim and Dry Creek to Upper Cambrian. Since the Gallatin formation, defined by Peale, includes both Middle and Upper Cambrian beds, Deiss (1936, p. 1341) pointed out that if the term Gallatin be retained it should refer to a group comprising only Upper Cambrian formations—the Pilgrim limestone and Dry Creek shale. Deiss' description of the Nixon Gulch section is applicable to the Cambrian in the area west of Three Forks, and, with two exceptions, thicknesses listed are approximately the same. The Meagher limestone apparently thickens westward, as it is 347 feet thick at Nixon Gulch, 430 feet about 4 miles northwest of Willow Creek village, and 420 feet on the west side of South

Boulder Creek, 7 miles south of Jefferson Island. The Pilgrim limestone, northwest of Willow Creek, however, is only 240 feet thick, compared with 402 feet at Nixon Gulch.

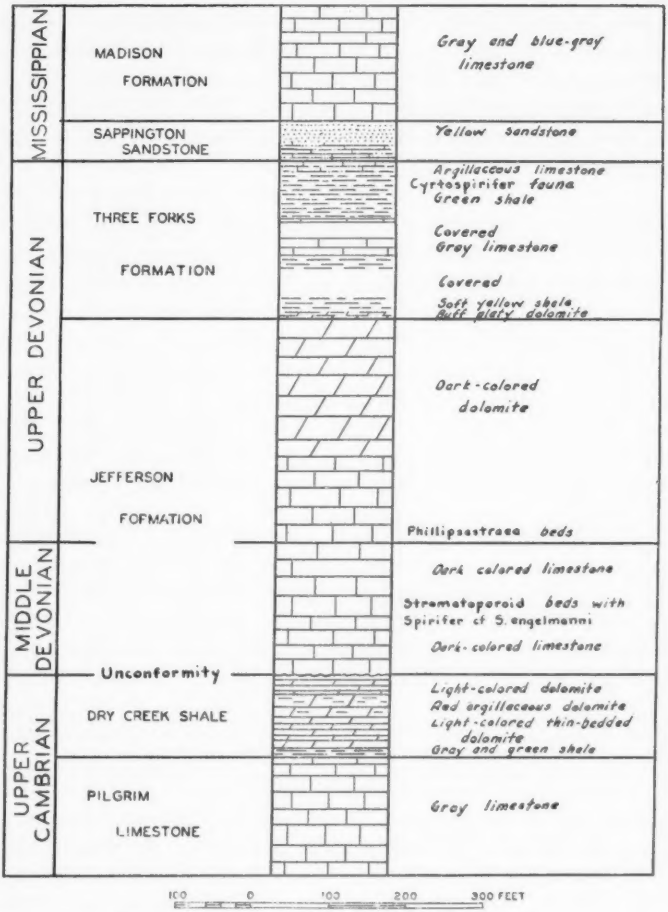

FIGURE 3.—*Middle part of the Paleozoic section*
One and a quarter miles northeast of Logan, Montana.

Red argillaceous dolomite occurs in the upper part of the Cambrian section (Fig. 3). A red color at this horizon may indicate unconformable relations with the underlying Upper Cambrian strata; and the writer was

unable to find any identifiable fossils in these beds or in the interval between them and the Jefferson limestone, in the lower part of which a Middle Devonian fauna is present. However, Deiss (1936, p. 1337; 1938, p. 61) reported Upper Cambrian trilobites and stated that "deposition of red, green, and buff mudstones and shales was characteristic of the shoaling phase of the Croixian sea." These beds are therefore included in the Dry Creek shale. The contact with the Jefferson at the top of the light-colored limestones and dolomites represents a hiatus involving the Ordovician, Silurian, and Lower Devonian, although no angular discordance is apparent.

DEVONIAN

Jefferson formation.—Since the Jefferson was originally described, uncertainty has existed regarding its age. In Montana and adjacent States to the south it has been variously assigned to the Devonian-Silurian, Devonian, Lower and Middle Devonian, and in recent reports to the Middle Devonian. A study of the type locality has yielded some critical evidence, and for this reason the formation is discussed at some length.

The type locality of the Jefferson formation is in the vicinity of Three Forks, Montana, where Peale (1893, p. 26) applied the name to beds above the Cambrian and beneath the Three Forks shales. He specified no exact location for a type section but stated that (1893, p. 27-28):

> "They are well developed in the hills a mile or two north of the East Gallatin River and on the main Gallatin above the Gallatin bridge of the Northern Pacific railroad, ... They are also well exposed in the hills on both sides of the Missouri just below the junction of the Three Forks and on both sides of the Jefferson a few miles above its mouth."

Peale stated that the formation is 640 feet thick and described it as follows (1893, p. 28):

> "On the Gallatin the beds . . . are brown and blackish in color, and microgranular in structure, due to their being very crystalline, and occur in alternations of rather massive beds of the very dark limestones, from 10 to 15 feet in thickness, with lighter colored, more laminated limestones in layers of 2 feet to 5 or 6 feet each."

Chemical analyses included in Peale's report show the rock to be dolomite.

The only recognizable fossils obtained by Peale were found about 30 feet below the top of the formation. Walcott identified the following (Peale, 1893, p. 29): *"Spirifera disjuncta," "Chonetes macrostriata," "Smithia,* sp. undetermined," *"Orthis,* sp. undetermined." Since these "were recognized to be the same as those occurring in the Three Forks" (Peale, 1896, p. 2), which is undoubtedly Devonian, the Jefferson formation was also assigned to this period. However, the possibility of Silurian age for some of the lower beds was suggested.

Stratigraphic position and a distinctive composition, color, and petroliferous odor have been the basis for identification of the Jefferson formation in Montana, northwestern Wyoming, northeastern Utah, and southern Idaho, because fossils are notably scarce and generally poorly preserved.

In the Little Belt Mountains Weed (1900, p. 288) obtained from the Jefferson two small collections of corals. In one collection, Schuchert identified *Diplophyllum caespitosum* Hall and regarded it as representing a Silurian horizon. Girty considered the other collection Devonian and identified the following: *Stromatopora* sp., *Pachyphyllum* (cf. *P. woodmani* H. & W.), and *Acervularia* sp. Weed does not state where in the section these fossils were collected.

Kindle (1908b) discussed the stratigraphy and fauna of the Jefferson formation. In the section at Logan, Montana, his collection was "limited to a small coral of undetermined species, *Schuchertella chemungensis* cf. *arctostriatus*, and fragments of fish teeth" (Kindle, 1908b, p. 9). In the Philipsburg quadrangle he studied two sections: one on Boulder Creek, 2½ miles northwest of Princeton, and the other on the east fork of Rock Creek, about 20 miles south of Princeton. From middle and upper beds of the Boulder Creek section Kindle (1908b, p. 10-11) identified 14 forms, among which are: "*Favosites* cf. *limitaris*", "*Productella* cf. *subaculeata*", "*Schuchertella chemungensis* var. *arctostriata*", "*Atrypa missouriensis*", "*Spirifer engelmanni*", "*Spirifer argentarius*", "*Spirifer utahensis*". From the lower 50 feet of the Jefferson in the Rock Creek section he obtained (1908b, p. 12): "*Schuchertella chemungensis* var. *arctostriata*", "*Strophedonta* cf. *macrostriata*", and "*Spirifer engelmanni*".

At Labarge Mountain, northeast of Viola post office, Wyoming, Kindle recognized the "Jefferson" by its stratigraphic position and its lithology. The only fossils found were "*Zaphrentis* and fragments of another undetermined coral". (Kindle, 1908b, p. 13.)

In the northern Wasatch Mountains, in the "Wasatch limestone", Kindle found a section 1000 to 1200 feet thick with lithology similar to the Jefferson of Montana. A collection from these beds, at Green Canyon, Utah, contained eight species, among which are the following (Kindle, 1908b, p. 16-17): "*Productella spinulicosta*", "*Camarotoechia* sp.", "*Spirifer argentarius*", "*Leiorhynchus utahensis*", "*Spirifer disjunctus* var. *animasensis*".

In the sections east of Cache Valley, Utah (Green and Logan Canyon), Kindle (1908a, p. 127-128) found a Silurian fauna below the Jefferson in a "light-colored magnesian limestone". One of the species of this fauna, "*Halysites catenulatus*", ranges upward into the "dark limestones", although (1908b, p. 17-18) "it does not occur in precisely the same strata

as the Devonian fossils, . . .". Regarding this occurrence Kindle stated (1908b, p. 24):

"The presence of this coral in the Jefferson limestone supplements in an important way the evidence afforded by the Devonian fossils as to the age of the lower portion of the Jefferson. As previously stated, the Devonian fauna contains none of the fossils commonly considered diagnostic of the Lower Devonian, but the occurrence in the lower part of the Jefferson of *Halysites catenulatus* affords stronger evidence of a horizon earlier than Middle Devonian than would even the most characteristic early Devonian species. If the validity of this evidence be conceded, then we must conclude that both the Lower and Middle Devonian are represented by the Jefferson limestone."

Richardson (1913, p. 411) listed eight species, identified by Kindle, from the Jefferson dolomite of Laketown Canyon, in the Randolph quadrangle of northern Utah. Among these the most diagnostic are: *"Favosites* cf. *limitaris"* and *"Spirifer engelmanni"*.

Tomlinson (1917, p. 128) listed *"Atrypa missouriensis"*, *"Actinostroma* sp.", *"Alveolites goldfussi"*, and *"Anplexus* cf. *hamiltoniae"* from a "Jefferson" locality in Yellowstone Park and another near Livingston, Montana.

In the Three Forks area, Haynes (1916, p. 278-279) found, within 25 feet of the base of the formation, corals "identified as *Favosites* cf. *limitaris* Rom., which is rather common in much of the Jefferson limestone". He considered these corals indicative of Devonian age of the entire formation.

From several localities in the Bayhorse quadrangle, Idaho, Kirk identified 10 different fossils from the Jefferson (Ross, 1934, p. 962). The most characteristic and abundant are, apparently: *Favosites* cf. *F. limitaris*, *Cladopora* sp., *Stromatopora* sp., *Syringopora* sp., *Diplophyllum* sp., and *Atrypa reticularis*.

In a collection from Ophir, Utah, obtained near the top of a formation questionably assigned to the Jefferson, Kirk identified the following (Gilluly, 1932, p. 21): *Cyathophyllum, Syringopora* (2 species), and *Spirifer* sp.

About the only evidence of pre-Middle Devonian age for the Jefferson is the occurrence of *"Halysites catenulatus"*, east of Cache Valley, Utah (Kindle, 1908b, p. 15-16). Merriam (1940) discussed this and stated that (p. 71):

"The occurrence of *Halysites* above or in association with any of the listed Jefferson forms cannot well be reconciled with the known distribution of Devonian and Silurian faunas in the central Nevada area, . . ."

Merriam further pointed out that (p. 70):

"The meager described and listed faunas from the Jefferson formation completely lack the elements of the Great Basin Lower Devonian *(Trematospira* fauna, *Spirifer kobehana* zone) as well as species of the Middle Devonian *Spirifer pinyonensis* zone."

and (p. 68):

"The insecure evidence at hand would indicate that the known Jefferson species of the type area are, with the possible exception of *Favosites* cf. *limitaris* and *Atrypa* cf.

missouriensis, of Devils Gate age, but older than the *Cyrtospirifer* zone of the Devils Gate, which last is equivalent to the greater part of the Three Forks formation overlying the Jefferson."

Study of the Jefferson in the type area has yielded further evidence of the Devils Gate age of the formation. About 1¼ miles northeast of Logan, Montana (Fig. 3), is a good exposure in a gully north of the electric power line, where the Jefferson is approximately 450 feet thick. The lower 280 feet is mostly limestone, although somewhat dolomitic, and the upper 170 feet is dolomite. The rocks are largely thick-bedded and massive and various shades of gray and brown. Dark grayish-brown weathered surfaces and a petroliferous odor are characteristic. Chert concretions are present but not abundant. The base of the Jefferson is drawn between the light-gray and cream-colored limestone and dolomite beds (assigned to the Dry Creek) and the overlying dark, massive limestone beds. The upper limit is placed at the top of the dark dolomite.

Fossils were collected at three horizons in this Jefferson section. The lowest, about 75 feet above the base, yielded the following fauna:

Stromatoporoids
Spirifer cf. *S. engelmanni* Meek
Atrypa sp.
Schuchertella cf. *S. chemungensis* Hall

Approximately 175 feet above the base the following were collected:

Phillipsastraea ("Pachyphyllum") sp.
Spirifer utahensis Meek
Spirifer sp.
Atrypa cf. *A. missouriensis* Miller
A. reticularis (Linn)
A. sp.
Schuchertella sp.
Chonetes sp.
Platyceras cf. *P. conicum* Hall
P. sp.

A collection made by Dr. C. W. Merriam about 50 feet below the base of the Three Forks formation included:

Cyathophyllum sp.
Cyrtospirifer sp.
Atrypa, 2 species
Productella sp.
Schuchertella sp.
Pelecypods (at least 2 species)
Trilobite, sp.

The few poorly preserved fossils collected in the Logan section are diagnostic enough to make possible correlation with the Devils Gate formation of Nevada. The most diagnostic fossil is the coral *Phillipsastraea ("Pachyphyllum")*, collected about 175 feet above the base of the Jefferson at Logan. In accordance with Merriam's (1940, p. 9, 59-60) division of the Devils Gate formation, beds containing this

form mark the base of Upper Devonian. The only diagnostic fossil recognized in the Logan section below the *Phillipsastraea* zone is *Spirifer* cf. *S. englemanni.* Kindle (1908b, p. 12) also reported *"Spirifer engelmanni"* from the lower 50 feet of the Jefferson in the Philipsburg quadrangle, Montana. In the Nevada section, it and *Spirifer utahensis* occur in beds which probably belong in the *Spirifer argentarius* zone (upper Middle Devonian).

There is evidence, then, that the lower portion of the Jefferson below the *Phillipsastraea* zone, about 170 feet, is upper Middle Devonian and equivalent to the *Spirifer argentarius* zone of Devils Gate. No fossils have been found at Logan or reported from elsewhere in Montana, to indicate an older age.

The upper portion, including the *Phillipsastraea* zone, is Upper Devonian.

Three Forks formation.—The Three Forks formation conformably overlies the Jefferson and is about 200 feet thick. Peale (1893, p. 29-32; 1896, p. 2), Raymond (1907; 1909; 1910), and Haynes (1916a, p. 280-283; 1916b) described the lithology and fauna, but they included in the Three Forks (Upper Devonian) beds which should probably be assigned to the Mississippian.

The writer examined Devonian sections north of the Gallatin River at Logan (Fig. 3) and also about 4 miles north of Willow Creek village (Fig. 4). At Logan, the argillaceous limestone and shale in the upper part of the Three Forks contain an abundant *Cyrtospirifer* fauna. In the section north of Willow Creek the same fossils are very numerous in the green shales underlying the argillaceous limestone. The most abundant species are:

Spirifer raymondi Haynes
Cyrtospirifer monticola (Haynes)
C. whitneyi (Hall)
Ambocoelia gregaria Hall
Leiorhynchus (several species)
Camarotoechia contracta Hall
Cleiothyridina devonica Raymond
Schizophoria striatula var. *australis* Kindle
Rhipidomella vanuxemi Hall
Productella, 2 species

CARBONIFEROUS

Sappington sandstone.—North of Willow Creek, about 5 feet of carbonaceous shale separates the green shales and the argillaceous limestone (30 feet thick). The latter is in turn overlain by 60 feet of yellow sandstone, which, in the lower part, is argillaceous and calcareous.

A "mingling" of Devonian and Carboniferous fossils in the upper part of the "Three Forks" was reported by Peale (1893, p. 32; 1896, p. 2).

Raymond (1907, p. 119) stated that the fauna of the "yellow sandstone zone" indicated a "transition into the Mississippian (Madison limestone)." Schuchert (1910, p. 546) examined Raymond's collection and commented as follows:

"Among the fossils *Syringothyris carteri* and *Spirifer* cf. *striatus* are prominent. This fanule is to be compared with that of the lower Louisiana limestone of Pike County, Missouri. Therefore there is here, as in the Mississippi valley, a break in deposition clearly distinguishing the Devonic, both physically and faunally, from the Mississippic."

Haynes (1916b, p. 27-28) studied the upper portion of the "Three Forks" in the field and failed to find "any indication of any unconformity in the section in this part." He noted, as did Raymond, that "*Spirifer whitneyi*" and other typical Upper Devonian forms present in the green shales and "nodular limestones" were not found in the overlying beds. However, Haynes did report that (1916b, p. 28):

". . . certain forms, such as *Rhipidomella vanuxemi* (?) and *Productella* cf. *arctostriatus* were sufficiently abundant in both the gray limestone, . . , and the yellow shale, . . , which almost immediately overlies (the gray limestone) at Logan, . . , to indicate that there is no sharp break in the record here."

Haynes agreed with Raymond that the fauna represents a transition from Upper Devonian to Mississippian.

North of Willow Creek (Fig. 4), the following fossils were collected at the base of the yellow sandstone:

Streptorhynchus (?) sp.
Productus cf. *P. arcuatus* Hall
Camarotoechia cf. *C. tuta* (Miller)
Spirifer cf. *S. marionensis* Shumard
Syringothyris cf. *S. hannibalensis* Swallow

No species of this *Syringothyris* fauna were found in the *Cyrtospirifer* zone, nor were any species of the latter fauna found in the yellow sandstone. However, since the two localities are stratigraphically 30 feet apart, the possibility of a transitional fauna in the intervening beds cannot be denied. No marked break in deposition is apparent from the top of the carbonaceous shale to the base of the Madison limestone. However, it seems probable that the lower part of the argillaceous limestone and the carbonaceous shale belong in the *Cyrtospirifer* zone, since this Upper Devonian fauna occurs in similar beds above the green shales of the Logan section. The base of the *Syringothyris* zone is drawn at the bottom of the yellow sandstone. There is a sharp contact between the sandstone and the overlying Madison, and the fauna of the Madison is distinctly different.

The yellow sandstone is both lithologically and faunally a recognizable stratigraphic unit. Since it is exposed at a number of localities within a few miles of the village of Sappington, the name Sappington

sandstone is proposed. At the type locality NW ¼ sec. 36, T. 2 N., R. 1 W.) the thickness is 60 feet.

According to Schuchert (1910, p. 546), the fauna of the Sappington is comparable with that of the lower Louisiana limestone of Pike County, Missouri. Moore (1928; 1935) assigned the Louisiana to the lower Mississippian and listed (1928) the faunas of this and other units comprising the Kinderhook group of northeastern Missouri. *Productus* cf. *arcuatus* is reported in the Hannibal formation, but no species of *Productus* are listed for the underlying Louisiana limestone or older beds. *Spirifer marionensis* is stated to be characteristic of the Louisiana (1928, p. 58), but it is listed also for the underlying Saverton shale and the overlying Hannibal formation. *Syringothyris hannibalensis* is reported from the Saverton shale and Louisiana limestone. These facts suggest that the Sappington sandstone is Kinderhook, but more specific correlation does not seem warranted.

Madison formation.—A thick section of Madison limestone conformably overlies the Sappington. Peale (1893, p. 33-39) described three lithologic divisions, with a total thickness of 1250 feet. These are, from the base upward: "Laminated limestones" (325 feet), "Massive limestones" (350 feet), and "Jaspery limestones" (500-600 feet). Although these members are readily recognizable in the field, the exact boundaries are not sharply defined in most places. The writer measured two sections of the Madison, one about 5 miles south of Jefferson Island, and the other (Fig. 4) 4 miles north of Willow Creek village. The thickness at both localities is approximately 1800 feet. The "Laminated limestones" and "Massive limestones" probably comprise the lower 1200 feet of the section, in which case they have a combined thickness nearly twice that which Peale assigned to them. The thickness of the "Jaspery limestones" agrees with Peale's figure of 500-600 feet.

West of Three Forks, the Madison formation (Fig. 4) includes a limestone conglomerate not mentioned in previous reports on this area. Although in most places obscured, the conglomerate was observed at three localities north of the Jefferson River. The best exposure is on the northwest side of a canyon about 4 miles north of Willow Creek, where the conglomerate is 50 feet thick and lies approximately 1200 feet above the base of the Madison. Poorly rounded, gray limestone pebbles and boulders are embedded in a matrix of soft, argillaceous and arenaceous, red limestone. Although angular discordance is not apparent, the conglomerate extends for at least 10 or 15 miles, indicating a significant stratigraphic break. Below it the limestone contains an abundant lower Mississippian fauna. Upward, the conglomerate grades rather

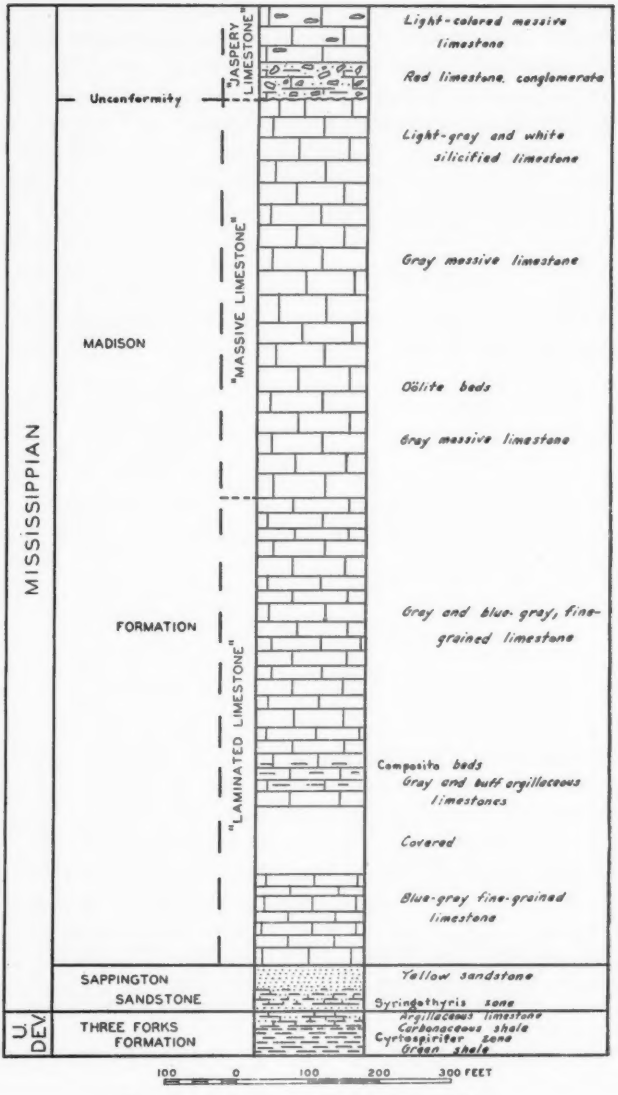

FIGURE 4.—*Devonian-Mississippian section*
Four miles north of Willow Creek village, Montana.

abruptly into beds which are probably Peale's "Jaspery limestones." In the upper strata, fossils are fewer and less well preserved, and the beds may be middle or upper Mississippian.

Four miles north of Willow Creek (Fig. 4) the following fossils were collected 280 feet above the base of the Madison:

Chonetes logani Norwood and Pratten
Linoproductus altonensis (Norwood and Pratten)
L. cf. *L. ovatus* (Hall)
Spirifer centronatus Weller
Hustedia sp.
Composita humilis
Eumetria verneuiliana (Hall)

A collection in the same section, but 625 feet above the base of the Madison, contained the following:

Caninia sp.
Michelinia sp.
Crinoids
Rhipidomella sp.
Schizophoria compacta Girty
Schuchertella cf. *S. chemungensis* Hall
Schellwienella sp.
Chonetes logani Norwood and Pratten
Dictyoclostus cf. *D. galeanus* (Girty)
Linoproductus cf. *L. ovatus* (Hall)
Spirifer centronatus Weller
Spiriferella sp.
Spiriferina solidirostris White
Rhynchotetra elongata Weller
Rhynchopora sp.
Eumetria perstrialis Rowley
Composita humilis
Platyceras sp.

Big Snowy group.—The Big Snowy group, according to Scott (1935, Fig. 3), lies between the Madison and Amsden in central Montana and is present only in the northeastern part of the Three Forks quadrangle. However, a few miles west of Three Forks, Scott figured (1935, Fig. 1) a thickness of about 30 feet for the Otter shale, which belongs to this group. The writer was unable to recognize this formation except in the hills north of Three Forks, where 3 feet of green shale, which may be Otter, separates the Madison and Amsden. Because the shale is thin and of limited extent it is not represented on the geologic map.

Amsden formation.—Peale included in the Quadrant formation beds which overlie the Madison and which are overlain by "the prominent bed of quartzite which has been provisionally taken as the base of the Mesozoic." In this interval he described two lithologic members: "Red limestones" (170-200 feet) and "Cherty limestones" (150-180 feet).

However, Scott (1935) showed that these beds are not equivalent to the Quadrant at the type locality in Yellowstone Park but underlie the true Quadrant and should be correlated with the Amsden of south-central Montana and north-central Wyoming.

Branson (1937) proposed the name Sacajawea formation for the lower part (Mississippian) of the Amsden section in Wyoming, in which Darton (1904, p. 396-397) included both Mississippian and Pennsylvanian beds. In the Three Forks area the lower part of the Amsden, which is probably equivalent to the Sacajawea, contains an upper Mississippian fauna comparable to that of the Brazer limestone of southeastern Idaho (Mansfield, 1927, p. 63-71). From beds in the lower part of the Amsden, 6 miles northwest of Three Forks (NE ¼ sec. 32, T. 3 N, R. 1 E.), the following species were obtained:

Linoproductus sp.
Streptorhynchus sp.
Spirifer sp.
S. pellaensis Weller
Composita cf. *C. sulcata* Weller

A collection from beds about 75 feet above the base of the Amsden (Fig. 5), 6 miles west of Three Forks, contains the following species:

Caninia sp.
Chonetes loganensis
Dictyoclostus sp.
"Gigantella" brazeriana (Girty)
Linoproductus cf. *L. ovatus* (Hall)
Spirifer cf. *S. pellaensis* Weller
Composita sp.

Scott (1935, p. 1023), after a study of the faunal and stratigraphic relations, concluded that "the Amsden of central Montana must be middle or upper Chester in age." However, the writer found beds in the upper 100 feet of the Amsden, 2 miles northwest of Sappington, in which a large species of *Linoproductus, Spirifer opimus* Hall, and *Composita* cf. *C. subtilita* (Hall) are very abundant. These fossils are generally considered Pennsylvanian.

Quadrant formation.—The "prominent bed of quartzite," referred by Peale to the base of the Mesozoic (Ellis), comprises the Quadrant formation as it is now understood. In this area the quartzite is about 150 feet thick.

Although fossils are scarce in the Quadrant, the known faunal and stratigraphic relations indicate Pennsylvanian age. From Jefferson Canyon, west of Three Forks, Scott (1935, p. 1020) reported a number of sponge spicules "analogous to those . . . from the Pennsylvanian

of Illinois and Indiana." Scott (1935, p. 1019) also stated that the Quadrant "is unquestionably a westward extension of the Tensleep sandstone."

Phosphoria formation.—According to Condit *et al.* (1928), the Phosphoria formation (Permian) overlies the Quadrant in the Three Forks region. Apparently, however, Scott did not agree, since in his columnar sections for this region (1935, Fig. 1) the Jurassic directly overlies the Quadrant. In the hills south of Summit Valley several thin phosphate beds seem to establish the presence of the Phosphoria. The vertical limits of the formation cannot be accurately drawn, but the thickness does not exceed 70 feet. However, in the northeastern part of the area, the basal sandstone of the Ellis is red and lies directly on the Quadrant. Apparently the Phosphoria, if deposited, was eroded before the invasion of the Jurassic sea.

JURASSIC

Ellis formation.—Approximately 170 feet of brown and gray sandstone, light-colored cherty dolomite, and chert comprise the Ellis formation. In the middle and upper parts, calcareous sandstones contain numerous poorly preserved marine fossils, which appear to be Jurassic.

CRETACEOUS

Kootenai formation.—Approximately 1500 feet of nonmarine strata overlies the Ellis, without apparent unconformity. Although two limestone members are present, sandstone and red and green shale comprise most of the section. In the upper part of the section there are several thin seams of lignite. The limestones contain numerous gastropod remains, and the sandstones near the top of the formation have poorly preserved plant fossils. Peale mapped the "gastropod limestone" and underlying beds as the Dakota formation, but they have since been correlated with the Kootenai of northern Montana and Alberta (Fisher, 1908). Peale mapped the lignite beds, which lie above the uppermost "gastropod limestone", as part of the Laramie and thus separated them from the Dakota by the "Montana and Colorado" (1800-2000 feet). East of Jefferson Canyon, the lignite lies stratigraphically less than 100 feet above the limestone and less than 300 feet below the base of the Livingston conglomerates. The lignite in Summit Valley lies beneath the Colorado shale. Therefore, the lignite is considered part of the Kootenai, and apparently is not represented in this area.

The lower part of the section, here included in the Kootenai, may be equivalent to the Morrison and Cloverly formations, which have been mapped in the Livingston quadrangle to the east. However, paleontologic evidence is lacking.

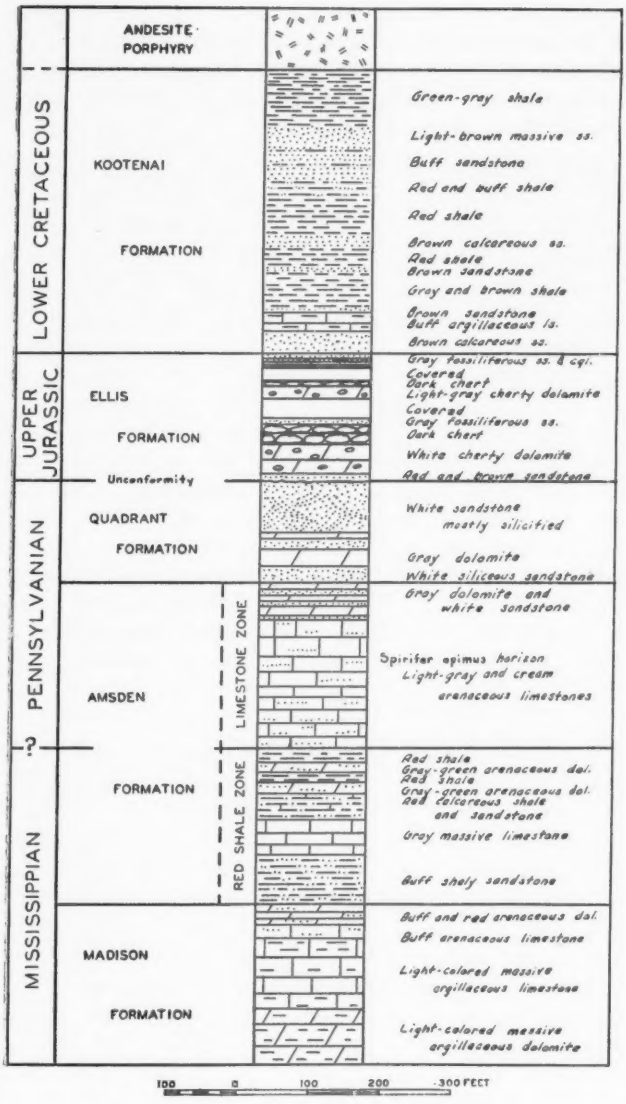

FIGURE 5.—*Upper Paleozoic-Mesozoic section*

Six miles west of Three Forks, Montana.

Colorado shale.—In Summit Valley, about 100 feet of black shale overlies the sandstones of the upper Kootenai. Elsewhere in the area the shale was apparently removed by pre-Livingston erosion. On the basis of lithology and stratigraphic position, the beds are mapped as the Colorado shale (Upper Cretaceous).

Livingston formation.—The volcanics of the Livingston formation are separated from the underlying Kootenai and Colorado by an erosional unconformity. Sandstones and a few thin beds of tuff are present, but the Livingston is composed predominantly of lavas and coarse conglomerates derived from them. Although basalt and rhyolite occur in the upper part of the section in South Boulder valley, most of the flows are andesites.

The stratigraphic position and structure of the Livingston indicate that it was formed during volcanic activity which accompanied the Laramide revolution. In the Three Forks area, it has been folded with the underlying strata, though apparently not so intensely, and must, therefore, have been formed before the deformation was completed. Probably the volcanics should be correlated with similar rocks south of Helena (Pardee and Schrader, 1933, p. 188), which are older than the Boulder batholith and probably "late Cretaceous". According to Stone and Calvert (1910), the Livingston formation in the area to the east is equivalent to the Claggett, Judith River, Bearpaw, and Lennep formations of Upper Cretaceous and to the Eocene Lance and lower Fort Union.

TERTIARY

Flat-lying beds of gravel, sand, clay, marl, and volcanic ash comprise the Tertiary deposits. They are almost entirely confined to the lower elevations and range in thickness from a few feet to possibly several hundred feet near Three Forks. Throughout most of its course, the Jefferson River flows in a broad open valley, underlain by Tertiary deposits and recent alluvium. However, at Jefferson Canyon and two localities to the east, the river has cut deep gorges in pre-Tertiary rocks. These gorges probably resulted from superposition of the stream from the once much thicker and more extensive Tertiary cover.

Peale mapped these Tertiary sediments as the "Bozeman lake beds" (Neocene). Douglass (1903, p. 151-155), Osborn (1909, Fig. 10; 1908, Fig. 5), and others studied a number of the vertebrate fossils and assigned them to Oligocene and Miocene. Douglass (1903, p. 154-155) applied the name "Madison Valley Beds" to Miocene deposits which form the top of the bluff on the east side of the Madison River, south of Logan.

IGNEOUS ROCKS

INTRUSIVES

Sills of syenite, trachyte, olivine trachyte, and trachyte porphyry are relatively common in the Three Forks area. They are all mapped as syenite. They intrude Cambrian to Cretaceous strata. Because the sills parallel the folds of the sedimentary rocks, they doubtlessly were intruded before or in the early stages of the Laramide deformation.

In the northern part of the area, several andesite porphyry stocks intrude the Cretaceous formations. North of Three Forks, a small stock with average composition of a quartz monzonite intrudes Paleozoic strata. The stock includes monzonite, quartz monzonite porphyry, and in the western part, quartz latite porphyry. South of Willow Creek, there is an intrusion of quartz latite and quartz latite porphyry; and another intrusion of similar composition on South Boulder Creek. At both the latter localities the intrusives are bounded by basement and Cambrian formations. With the possible exceptions of these last two, the stocklike masses appear to have been emplaced after the deformation of the sedimentary section and before deposition of the Tertiary "lake beds".

EXTRUSIVES

Most of the extrusive rocks in this area are included in the Livingston formation. Although flows of andesite and andesite porphyry are predominant, basalt and rhyolite occur in the upper part of the section along South Boulder Creek.

Small areas of basalt on Antelope Creek and on the north side of Jefferson River, between Willow Creek and Sappington, seem to be Upper Tertiary extrusives. The relations are not proved. The basalt appears to overlie the "lake beds" north of Jefferson River, in which case it is Miocene or younger.

STRUCTURAL GEOLOGY

GENERAL FEATURES

The pre-Cambrian rocks of the Tobacco Root Mountains show two periods of strong deformation preceding deposition of the Belt sediments (Tansley and Schafer, 1933, p. 22). The record indicates no major structural adjustments in the Three Forks region during the Paleozoic and Upper Cretaceous. In the Upper Cretaceous and early Tertiary, however, the strata were strongly deformed during the Laramide revolution. The Tertiary "lake beds" and valley alluvium are the only sediments not appreciably disturbed.

Laramide deformation produced several complex structural trends (Fig. 1). South of Jefferson River, the axes of folds, and the associated

thrust faults, strike northwest. To the north and northwest, along the river and in the area southwest of Jefferson Island, the general structural trend is N. 60-70° E. In the northern half of the region, and along the west boundary, south of Jefferson River, the trend is N. 15-35° E., swinging to approximately north along the northern limit of the area.

FOLDS

In the north, two anticlines just west of the andesite intrusion are overturned toward the west. All other folds north of the Jefferson River are overturned toward either the east or the southeast (Pl. 1). Dips on the west flanks of the anticlines average about 40°, the east and southeast limbs being in most places nearly vertical. Three anticlines south of the Jefferson River are overturned to the southwest.

The largest folds were probably controlled directly by deformation of the basement. However, the Madison limestone, which lies near the middle of the stratigraphic section, was also a factor in development of the symmetry and other characteristics of the folding. For instance, some degree of concentric folding, with the Madison as a center of curvature, is indicated by the number of anticlines which decrease in magnitude downward from the Madison, and the several synclines which decrease upward toward it. Isoclinal folding is confined mostly to the Cambrian formations, where they are underlain by the Belt series.

FAULTS

Thrust faults.—Nearly all the faults exposed in the Three Forks area are high-angle thrusts. Low-angle thrusts with displacements of more than a few feet were not observed. Most of the high-angle faults are associated with folds and resulted from shearing of the beds on the steep limbs. The faults cut across the folds only in the area of Jefferson Canyon and southwest of Jefferson Island.

Lombard overthrust.—In the canyon of the Missouri River, near Lombard, about 15 miles north of Three Forks, the Lombard overthrust (Haynes, 1916a, p. 271-272) dips about 40° W. The stratigraphic throw is "about 6,800 feet", but the total displacement is "approximately two miles". Pardee and Schrader (1933, Pl. 1) suggested that to the north the Lombard connected beneath the Tertiary cover with the El Dorado overthrust, northeast of Helena. The fault trends southwestward from Lombard and disappears beneath undisturbed Tertiary sediments just north of Three Forks. There the fault dips 50-60° W., and the throw is about 2500 feet.

Jefferson Canyon fault.—North of Sappington there is an east-west, high-angle thrust with an apparent stratigraphic throw of approximately

12,000 feet. This thrust is named the Jefferson Canyon fault. Belt arkose and the basement gneiss were thrust against volcanic conglomerates of the Livingston formation. At Jefferson Canyon, a cross fault offsets the major fault northward and apparently is in part responsible for a sharp change in trend to about S. 70° W. The stratigraphic throw is decreased at this locality to approximately 6000 feet, the Belt having been brought against the folded Madison. Southwest of Jefferson Island, beyond a cover of Tertiary "lake beds", the Jefferson Canyon fault reappears, and the stratigraphic throw is here 3000 feet. The apparent variation in the magnitude of the Jefferson Canyon fault may be accounted for, in part, by the distribution of the displacement among several, associated, smaller faults. The aggregate stratigraphic throw in the Jefferson Canyon section is at least 8000 feet. The thrust also appears to have been distributed along two faults southwest of Jefferson Island, where the combined displacement is at least 6000 feet. Major folding and steep plunges are responsible partly for large variations in the apparent displacement along the fault.

This zone of thrusting delimits the northwest structural trends and coincides with the southern limit of the Belt sediments. Probably the Belt shore line was relatively steep. Apparently this declivity in the basement surface was an important factor in localizing the zone of thrusting. Indeed, the Belt shore line itself probably overlay a zone of weakness in the gneiss and this really controlled the development of the Jefferson Canyon fault.

Normal faults.—Two normal faults are shown on the areal map. One, south of Sappington, involves the basement and the Cambrian formations. The maximum displacement is not more than 800 feet and decreases rapidly along the strike. The other fault is between the Livingston and Madison formations, in the northwestern part of the area. Projections of folds north of Three Forks, and especially the folds in the Boulder valley, indicate that normal faults are present in the rocks beneath the Tertiary "lake beds." Since the "lake beds" in these areas are relatively undisturbed, this faulting must be pre-Oligocene. In the Missouri valley, south of Townsend, however, Pardee (1925, p. 32) reported normal faults, with displacements of 50-150 feet, in the Oligocene and Miocene strata.

RELATION TO REGIONAL STRUCTURES

North of the Jefferson Canyon fault, the structures conform to those of the Northern Rockies, both in their general orientation and in the plainsward thrusting and overturning (Fig. 1). The more easterly trends through the Jefferson Canyon region have apparently been controlled by a zone of weakness in the basement and probably represent only a local

variation in the more general north-south trends of the Northern Rockies. In this connection, Chamberlin (1941) pointed out that in the Rocky Mountains, where the basement rocks were deeply buried beneath the Paleozoic and younger sediments, the Laramide deformation resulted in relatively regular, long, linear, parallel ranges; but in the area to the east, where the sedimentary section was thinner, basement control upset the regularity of the deformation and resulted in variously oriented structures.

South of the Jefferson Canyon region, the folds, and associated thrust faults, strike northwest, paralleling the structural trends of the pre-Cambrian gneiss in the northern Tobacco Root Mountains. This northwest structural trend has also been recognized in the northern Madison and Gallatin ranges and in the Beartooth uplift. The change from the usual north-south trend is approximately marked by a line through Butte, Whitehall, Bozeman, and Livingston.

Although all the major structures involving the sedimentary section were developed during the Laramide revolution, the trends other than generally north-south may well have been initiated earlier. Thom (1923, p. 12) cited structural and stratigraphic evidence indicating ". . . that uplifts, probably of east-northeast orientation, were developed in central Montana at various times during the Mesozoic and later Paleozoic, . . ." In fact, trends other than north-south are relatively common in the Rocky Mountains, other instances being the Little Belt-Big Snowy anticlinorium of Montana, the Wind River basin and Sweetwater anticline of Wyoming, and the Uinta Mountains of Utah.

Available information regarding west and northwest trends in southern Montana indicates that overturning and thrusting to the south and southwest is characteristic. In the Three Forks region, south of the Jefferson Canyon fault, three anticlines are overturned southwest, and associated high-angle thrust faults dip northeast. In the northern part of the Madison Range, Peale (1896) mapped northwestward-trending faults, and although apparently not certain of the relationships, he showed them in cross sections as high-angle northeast-dipping thrusts. In the Trail Creek-Canyon Mountain area, southwest of Livingston, Skeels (1939) mapped northeastward-dipping thrusts, and folds which are overturned to the southwest. The thrusts are similar in trend and direction of relative movement to those of the adjacent northern flank of the Beartooth uplift (Bucher, Thom, and Chamberlin, 1934, p. 173-174). Skeels (1939) explained the Trail Creek-Canyon Mountain structures as the result of resolution of northeastward Laramide stresses into components normal to the Bridger and Beartooth "blocks". However, the total area in which thrusting and overturning to the south and southwest has taken place is extensive. Thus more than local resolution of east-northeastward stresses

is indicated. Unfortunately, detailed information regarding Rocky Mountain structure in southern Montana is generally lacking, and it therefore seems unwise to draw any further conclusions regarding the origin and mechanics of the structures in the Three Forks region. That they are typical of Rocky Mountain structure over a wide area in southern Montana appears to be demonstrated.

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c/o Charles Nevin, Cornell University, Ithaca, New York.
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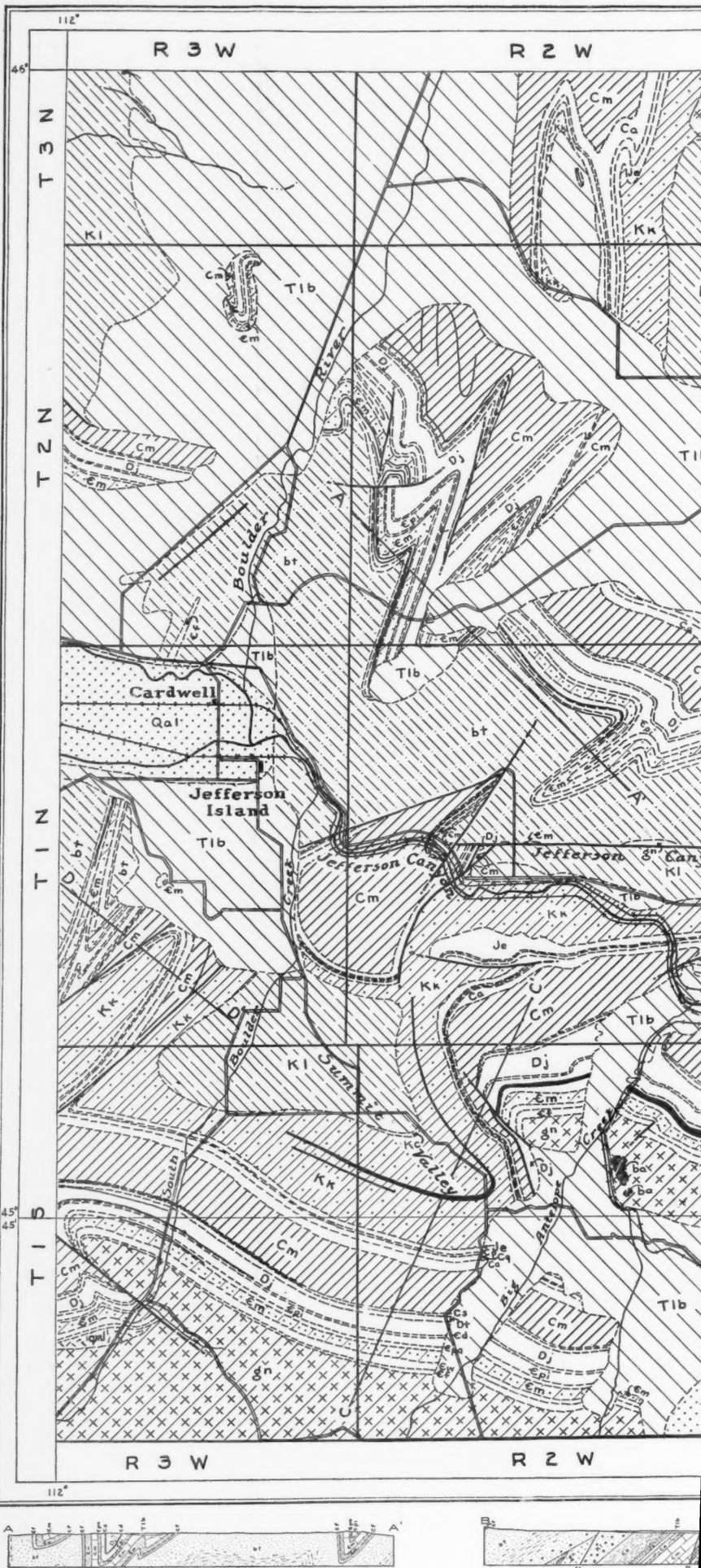
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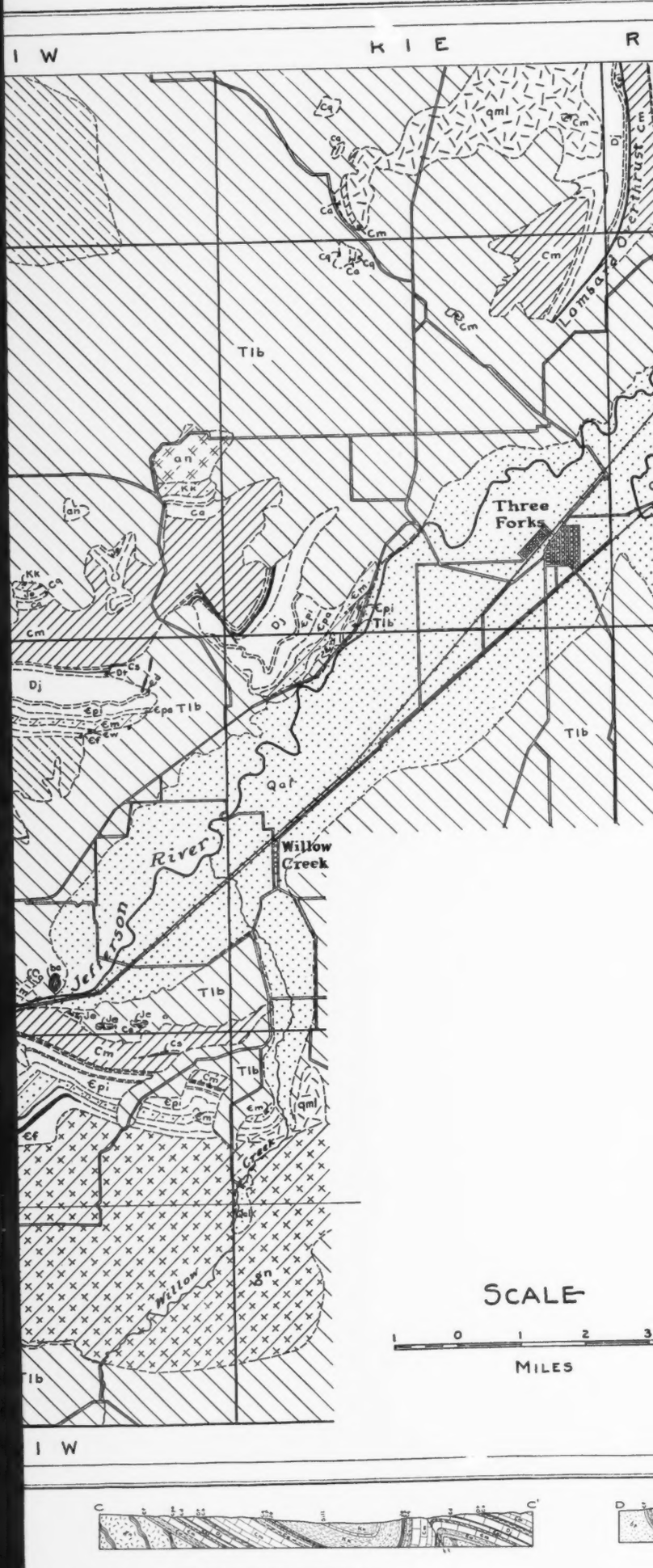
GEOLOGIC MAP AND STRUCTURE SECTIONS, THREE FORKS, MONTANA



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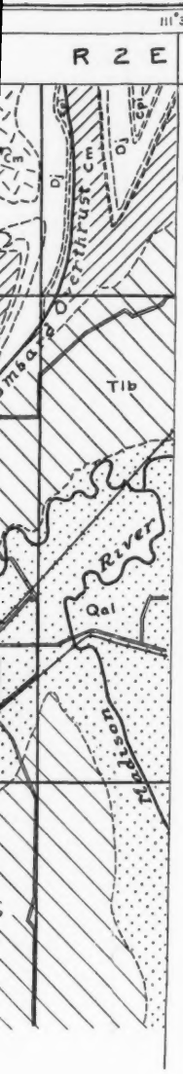

EXPLANATION

QUAT.	Qal	Alluvium
TER.	Tlb	"Lake Beds"
CRETACEOUS	Kl	Livingstone fm.
	Kc	Colorado sh.
	Kk	Kootenai fm.
JUR.	Je	Ellis fm.
CARBONIFEROUS	Cp	Phosphoria fm.
	Cq	Quadrant fm.
	Ca	Amsden fm.
	Cm	Madison fm.
	Cs	Sappington ss.
DEVONIAN	Dt	Three Forks fm.
	Dj	Jefferson fm.
CAMBRIAN	€d	Dry Creek sh.
	€pi	Pilgrim ls.
	€pa	Park sh.
	€m	Meagher ls.
	€w	Wolsey sh.
	€f	Flathead ss.
	bt	Belt Series
	gn	Gneiss
	ba	Basalt
	an	Andesite
	qml	Quartz-monzonite Quartz-latite
		Syenite
		Fault

46°
45°
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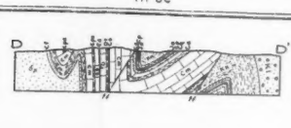

STRUCTURE SECTIONS
THREE FORKS, MONTANA
SCALE
MILES

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KEEWATIN END MORAINES IN ALBERTA, CANADA

BY J HARLEN BRETZ

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ABSTRACT

The outermost Keewatin moraines, within sight of the Rocky Mountains, are as youthful topographically as is the Late Wisconsin Altamont terminal moraine in eastern Alberta and western Saskatchewan. Differences in outline of the ice front, however, indicate that all moraines west of the Altamont are the product of an earlier advance. The moraine pattern records a southern Alberta convexity or lobe, limited on the west by the foothills of the Rocky Mountains and on the south by the Arctic Ocean-Gulf of Mexico divide. A complicated succession of melt-water dischargeways and associated glacial lakes is intimately related to the moraine succession. Escape of all this drainage was across the divide and to the Missouri River in Montana. Correlation of discontinuous moraine units is possible because drainage held close to the edge of the Keewatin ice sheet as it retreated down the slope of the Great Plains. Moraine outside (higher than) any one drainage channel thus is older than moraine inside (lower than) the channel.

INTRODUCTION AND ACKNOWLEDGMENTS

A study of the age and distribution of Albertan end moraines, urged by Wellington D. Jones, was aided by a grant from the Penrose Bequest of The Geological Society of America. When Jones became unable to do field work the writer, assisted by Paul Herbert, undertook the study. One aim of the project was left unrealizable by Jones' enforced withdrawal—namely, a study of soil profiles to aid in determining the relative ages of drift sheets in the region.

During 50 days of field study, 9000 miles of traverses were made by automobile, and the moraines of approximately 40,000 square miles

of the plains of Alberta were mapped. This was preceded by a week of field study of moraines previously mapped in Montana, Saskatchewan, and extreme southern Alberta. Throughout this preliminary study the writer found himself in almost complete agreement with Alden of the U. S. Geological Survey and with Johnston and Wickenden of the Geological Survey of Canada. Neglecting a few slight changes in boundaries, the Saskatchewan moraines were then traced northwestward into Alberta to Lat. 54° N., where forest, muskeg, and few and execrable roads forbade the detailed mapping practiced in the bald prairies and copse-dotted country farther south. In that mapping no continuous moraine boundary was drawn on any unviewed tract more than 5 miles across.

No fixed rules can be laid down as to what constitute moraine boundaries. End moraines with only mild relief may contrast as much with their adjoining ground moraine as the most strongly expressed end moraine where bordered by ground moraine of unusually marked relief. Thus ground moraine in one tract may have stronger expression than end moraine in another. If an area enclosed by moraine boundaries is to be considered a marginal thickening of drift, it should also have linearity of outline and subparallelism of the longer boundaries, but it is not necessarily a continuous belt. Much Albertan end moraine occurs in separate linear units, the elongations of which commonly point to other units, similarly oriented and elongated, a group constituting a discontinuous end moraine. Several successive and close-set rows of discontinuous moraines, with large interruptions, may offer difficulties in correlation.

No one moraine character can be expected to continue throughout miles of length. End moraines may widen and narrow, may be interrupted, or may change along their length in altitude, local relief, steepness of slopes, abundance of undrained depressions, and composition of till. Even the soil profile developed in postmoraine time changes with latitude in all the long north-south moraines of Alberta. Thus, in the absence of continuity, there is need for some other criterion for correlation. The plains of southern Alberta possess the needed criterion, a remarkably developed succession of marginal melt-water drainage courses.

Grateful acknowledgment is made for a grant from the Penrose Bequest of The Geological Society of America and for one from the Salisbury Fund of the University of Chicago, which together covered field and office costs of the investigation. Appreciation is expressed for the courtesy of the Geological Survey of Canada in supplying unpublished maps of glacial features in the region studied. Parts of these are in-

corporated in the map accompanying this report, on which the boundaries by Johnston and Wickenden of the Geological Survey of Canada are indicated. The writer is especially indebted to Wellington D. Jones of the University of Chicago for many valuable suggestions made before the field work was undertaken, for criticisms and comments during the ensuing office map work, and for a critical reading of the manuscript. Paul Herbert's unflagging and intelligently directed zeal was of inestimable value in the field and office work.

GENERAL FEATURES OF THE MORAINE PATTERN OF ALBERTA

Moraines shown by the earlier mapping in Montana and southern Alberta (Alden, 1932; Johnston and Wickenden, unpublished map) suggest a double lobate outline of the retreating ice front, the separating re-entrant coinciding with the present east-west divide between Hudson Bay and the Gulf of Mexico. Alden recognizes the southern lobation, naming it the Great Falls lobe, but his map shows only an interrupted terminal moraine, almost no recessional moraines. The existence of a northern or Albertan lobe is sufficiently well shown by Johnston and Wickenden though the text of their paper (1931) contains no specific mention of it. Many recessional end moraines of this lobe are shown in their later manuscript map.

Alden is definitely convinced that the Great Falls lobe recorded by the Montana moraines is "Early Wisconsin" in age. South of it lies an older drift sheet, Iowan or Illinoian in age according to Alden. Johnston and Wickenden are inclined to call their Albertan moraines "Early Wisconsin".

Warren (1937) described what he named the Viking moraine, his map showing the frontal edge to extend from the Cypress Hills northward to Lat. 54° 30′. In its southern part it crosses almost at right angles the moraine courses shown by Johnston and Wickenden. The scale of Warren's map is approximately 90 miles to the inch, so that the moraine course is much generalized. However, when superimposed, with the most liberal interpretation, on the more detailed mapping of Johnston and Wickenden and of the writer, it fails along more than three fourths of its total length to coincide with any end moraine units, and where it does coincide it involves portions of more than half a dozen successive recessional moraines as interpreted by Johnston and Wickenden, and the writer. Warren considered his Viking to be of Iowan or Illinoian age.

Jones (1941) published an abstract, without a map, demurring from Warren's conclusion regarding the age of the Viking and from his further suggestion that two older Keewatin moraines, west of the Viking,

may be Kansan and Nebraskan in age. Jones thought that even the westernmost of these moraines was " 'Young Drift' Wisconsin."

Rutherford (1941) published a map (scale about 50 miles to the inch) showing: (1) the eastern boundary of the westernmost of these moraines (the Duffield); (2) complete boundaries between parallels 52° and 55° of the next moraine to the east (the Buffalo Lake); (3) a course for the frontal edge of the Viking moraine which departs locally between 5 and 10 miles from the course shown by Warren; and (4) a frontal edge for the Altamont moraine (terminal Late Wisconsin) from the northwest corner of North Dakota diagonally across Saskatchewan and nearly to Lat. 55° in Alberta. He suggested that "the Viking moraine may represent an early stage in the ice advance that formed the Altamont." The Altamont moraine is accepted in the present study as terminal Late Wisconsin, equivalent to the Des Moines, Mankato, and Port Huron moraines. No attempt is made to correlate older moraines of the Great Plains region with the Early and Middle Wisconsin moraine succession in the Mississippi and Great Lakes regions.

Though there are four Keewatin moraines in Alberta in the general latitude of Edmonton, as stated in the three papers (Warren, 1937; Jones, 1941; Rutherford, 1941), and though they can be traced southward and southeastward and are successively younger from west to east, the writer's mapping and interpretations differ from those of Warren and Rutherford. The moraine patterns clearly indicate the existence of a lobe in southern Alberta; the pattern of the successively used and abandoned marginal glacial rivers proves it. Retreatal stages of this ice front were recognized by Dowling (1917) and by Williams (1929) when they outlined glacial drainage successions for far southern Alberta. No continuous ice front along the course of Warren's Viking can be made to fit into the details of this succession of drainageways. Large tracts shown on both Warren's and Rutherford's maps as having end moraine lack such features or other relief that could be mistaken for moraine. The Duffield, westernmost and therefore oldest of the Keewatin moraines in Lat. 53° 30′, has as marked a development of steep moraine hills and of steep-sided undrained hollows among them as has the Altamont. It has no more hillside gullying or notching of basins or alluvial fan encroachment into basins than has the Altamont. Topographically, both moraines are very youthful, comparable to Wisconsin moraines of the Great Lakes region in the United States. In contrast, the older Keewatin drift of Montana (called Iowan-Illinoian by Alden) rarely shows accentuated morainic topography and lacks undrained hollows almost entirely. Sites of

former hollows commonly are marked by filled flats, less commonly by encroaching gully heads.

Boulder-clad knolls and hills occur in all the moraines except possibly the Duffield. Road cuts through such hills reveal extraordinarily coarse drift in the upper few feet; boulders and cobbles are as abundant as on the surface. Many closely associated hills of similar size and form lack the boulder armor. The unusual composition appears due to melt-water washing of some knoll summits under the edge of the ice, not to subsequent slope wash as Warren argued.

Keewatin till in Alberta varies with locality, most of it having a high percentage of clay but some characterized by abundant silt and sand, with very little clay. Such variations seem due more likely to different Cretaceous and Tertiary source rocks overridden on the plains than to differences in amount of washing at time of deposition.

Indeed, the Alberta moraines possess surprisingly little accompanying outwash. Kame topography is rare and usually well localized. Eskers, valley trains, and glacial-lake deltas are equally rare. Many of the glacial rivers, flowing on low gradients, eroded channels 100 feet deep, hence could hardly have been heavily loaded with outwash sand and gravel. Widths of the channels suggest no excessive volume, no more perhaps than the present run-off across the plains. Johnston and Wickenden (1931) suggest, with reason, that wastage of the Albertan glacial ice may have been largely by evaporation.

A morainic feature suggesting this same idea is the relative steepness of the clay till piles. A water-saturated clay till deposited by a rapidly melting glacier may be expected to slide and settle under its own weight into low slopes. Thwaites (1939) visualizes the extreme of "seas of glacial mud," the till flattening out to constitute smooth drift plains. By this view, the clay till of these accentuated Albertan hills could not at time of deposition or later have contained any noteworthy amount of interstitial water.[1] Many sight-dip readings, made in tracts where average slopes were unusually steep and including nothing less than 10°, gave average clay till slopes between 14° and 17° and maximum slopes of 25°. Suspected kames were not measured.

Many of Alberta's canyonlike river valleys are clearly premoraine; the moraines are easily traced from the uplands down the valley walls, and some of them even part way across the valley bottoms. Battle River Valley near Hardisty shows this well. The valley of the Red

[1] Before this concept of dry deposition of the clay till hills can be accepted, we must consider that, when some moraines were built, the ice front was a glacial lake dam, and that other moraines became submerged in or surrounded by frontal lakes after the ice retreated from them. Should they not be considerably slumped and softened in expression by such experiences? Too few data regarding the topography of such moraines are at hand to justify any general statement.

Deer River has very well shown moraine crossings at Dorothy, at Hunting Hill north of Millicent, a few miles upstream from the Jenner ferry, a few miles downstream from the Buffalo bridge, and elsewhere. There has been little or no deepening of these valleys since the moraines were built.

There are also canyon-like coulees, now containing swamps and lakes, similarly crossed by moraines. Tramping Lake valley, in Saskatchewan, near the Alberta line, is an excellent example. Obviously the valleys of premoraine streams, they testify also that the present youthful valley system of southern and central Alberta was well advanced before the Wisconsin glaciation occurred. Their unfilled lakes and swamps tell as clearly as do the undrained moraine depressions that postglacial streams and slope wash have hardly begun to modify the glacial records.

Prevailing, Albertan morainic hills are nearly isolated mammillary forms of subequal sizes and proportions in any one locality. Elongated belts of such hills are far more common than are moraine ridges. Ice-block kettle holes among the hills are rare, though shallow undrained depressions abound. Exceptions to this rule are parts of the Duffield and the Altamont where hills are compounded by irregular piling, the most pronounced of these apparently being kame piles with associated ice-block depressions.

The eastern part of the Duffield, where crossed by the Edmonton-Jasper highway and its structure exposed in many fresh cuts, differs from any other moraine seen in this study. For a width of 8 miles it is composed (at least to the bottom of all cuts) almost wholly of silt, a considerable proportion of which is stratified. The bedding is lensy and somewhat irregular. Locally there is good cross-bedding, with fore-sets dipping westward. Some parts are sandy, a few lenses of pebbles occur, boulders are very rare. There is no deformation by ice thrust. Hollows and kettle holes appear to be molds of ice blocks. If the hilltops were flattish and accordant in altitude, this part of the Duffield would pass for a pitted plain. It stands as an accumulation of thousands of variously shaped and sized hills, rising nearly 100 feet above a ground-moraine plain to the east, and some large lake- and swamp-filled lowlands to the west. The Keewatin ice sheet doubtless lay immediately to the east when this moraine was deposited, but the cause of its western front is not clear. It was not Cordilleran ice for Keewatin moraine extends 20 miles farther west and Keewatin drift materials are known for 50 miles still farther. Residual Keewatin ice masses might have been present. Standing water is a possibility though this western front for at least 12 miles shows no deltaic forms.

FIGURE 1. SUBEQUAL MAMMILLARY HILLS
Typical of most Albertan Keewatin moraines. Buffalo Lake moraine near Mudspring Lake, Red Deer sectional map, Alberta.

FIGURE 2. TOPOGRAPHY CHARACTERISTIC OF THE DUFFIELD MORAINE
Near Rochfort Bridge, St. Ann sectional map, Alberta.

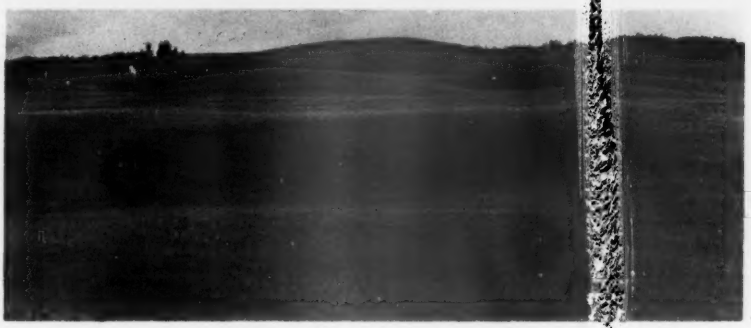

FIGURE 3. COMPOUNDED MORAINIC HILLS CHARACTERISTIC OF THE ALTAMONT MORAINE
East of Two Hills, Vermilion sectional map, Alberta.

ALBERTAN MORAINE TOPOGRAPHY

FIGURE 1. STREAM SCULPTURE ON SOUTH SLOPE OF NOSE HILL
The north slope of this hill, covered with moraine, possesses no such features, is much like that shown in Figure 2, this plate. Sullivan Lake sectional map, Alberta.

FIGURE 2. HILL WHOSE STREAM-CARVED TOPOGRAPHY HAS BEEN ALMOST OBLITERATED BY MORAINE
Some such hills have been erroneously interpreted as composed wholly of glacial drift. Sullivan Lake sectional map, Alberta.

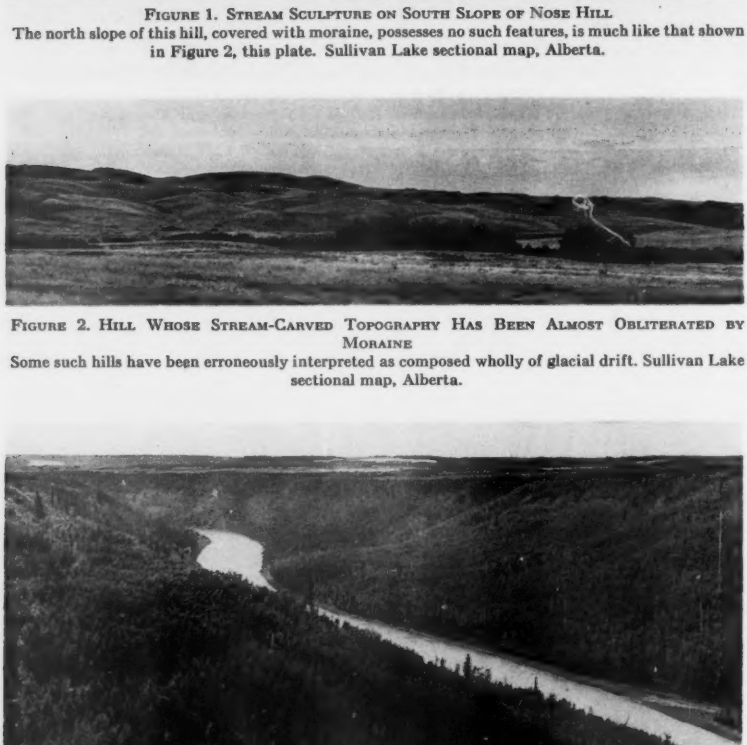

FIGURE 3. GORGE OF RED DEER RIVER NEAR BURBANK
Looking southeast from the divide transection. Summit of river bluff is a valley floor of the Albertan preglacial oldland. Hills of Delburne upland in the distance. Red Deer sectional map, Alberta.

West of this silt moraine along the Edmonton-Jasper highway, the Duffield moraine is less marked topographically but has very marked ice-thrust structures, with masses of incorporated shale and "Saskatchewan gravel" and with the parent underlying formations also deformed. Erdtman and Lewis (1931) have described some of these structures. The writer saw, in new road cuts, anticlinal overturnings of quartzite gravel in the basal till in which as much as 20 per cent of the pebbles were fractured as in folded conglomerates. Rutherford (1937) has reported similar fracturing in "Saskatchewan gravels" at a locality a few miles farther east and feels, as does the writer, that "this crushing may have been caused by glacial load or movement." It may well be that the Duffield as mapped is compounded of two moraines, the silt moraine recording a retreat of the ice edge and a readvance under sufficiently different conditions to produce the different composition, structure, and topography.

CORRELATION OF THE MORAINES

Because the Keewatin ice sheet retreated almost directly down the regional slope of the glaciated Great Plains, melt-water rivers prevailingly hugged the edge of the ice sheet, necessarily flowing essentially along the strike of the Plains, either toward the Gulf of Mexico or the Arctic Ocean. Thus as the ice edge retreated, newer and lower channels were formed as older and higher ones were abandoned. The same principle applies to a succession of ice-dammed lakes and pondings where a lack of continuous gradient along ice edges made rivers impossible. Moraine and premoraine elevations on the Plains make irregularities in the scheme but do not affect the principle.

A continuous marginal or submarginal glacial river course is a date marker for associated moraines. Those outside (higher than) the river course are older, those inside (lower) are contemporaneous or younger. Fortunately for our purpose, few of these ancient river courses functioned very long after the ice dam had withdrawn to a lower position.

Two assumptions underlie the use of this principle in the present correlation: (1) No postmoraine differential earth movements have occurred in sufficient magnitude to alter the relative altitudes of the glacial lakes and drainage ways on which this correlation is based. (2) No glacial water drained back under or over or through the various ice dams involved at different times during the glacial retreat.

Far to the southwest in the plains of Alberta occurred the earliest and highest stage of ponding and submarginal glacial stream discharge of this study, closely associated with the lower tract crossed by the St. Mary, Belly, Waterton, and Oldman rivers and with Milk River

Ridge to the southeast of it. As far as now known, there were two rivers and two lakes in the drainage line, Lake Cardston draining at 3900 feet above sea level [2] across a col in Milk River Ridge and down Lonely Valley to Milk River Lake which in turn discharged at the International Boundary across a 3450-foot col to a glacial river mapped by Alden from Coutts and Sweetgrass to and a little beyond Shelby, Montana. The two river courses were mapped by Johnston and Wickenden and by Alden respectively, Milk River Lake's outline is largely after Johnston and Wickenden, but the outline of Lake Cardston (drawn, as are those of all glacial lakes on the map, by tracing contours on the ice-free land of the stage at the outlet head's altitude) is a departure from Johnston and Wickenden's depiction of their Lake Magrath.[3] No glacial lakes of this study are recorded, so far as known, by shore lines.

The moraine recording the position of the ice edge at the Lake Cardston-Milk River Lake stage is fragmentary. Furthermore, it may not coincide with the latest and greatest spread of Lake Cardston. The ice may have withdrawn some miles from the northwestern part of the moraine, yet still have remained up against the north side of Milk River Ridge at 3900 feet or above. This qualification is to be applied in several later cases since a glacial lake must widen, as the ice retreats, from its initiation until its disappearance.

The north slope of Milk River Ridge is steep as Albertan plain surfaces go, and ice withdrawal of a few miles down this slope opened Middle Coulee Route, its head at 3350 feet, the lake it drained lying to the west in the same lowland as did its predecessor, Lake Cardston, but 550 feet lower. A few miles more of retreat opened Kipp Coulee route at 3250 feet. Because of extensive overlap of the Middle and Kipp lake stages with earlier Lake Cardston and later Lake MacLeod, depiction of the lake surfaces is omitted from the map (Pl. 3). These may be styled the Lake Magrath stages.

The discharge carried by both Middle and Kipp coulees entered Verdigris Coulee and flowed thence to Milk River a few miles east of the site of the earlier, higher Milk River Lake. Ice-front positions for these two dischargeways are almost without moraine record, but it is obvious where ice must have stood when each was active.

The main Verdigris Coulee discharge began after a very few miles of further retreat down this northern slope of Milk River Ridge. The head of Verdigris Coulee is a little below 3200 feet, its depth at

[2] The sectional maps of Alberta and Saskatchewan, none of which has a smaller contour interval than 50 feet, constitute almost the only source of altitude data for this study. Text or map figures ending in 00 or 50 indicate the value of the contour nearest to the feature being considered.

[3] Shown as discharging over the 3900-foot col. The locality of Magrath was under glacial ice at this stage.

the head too slight to be shown by the 50-foot contour interval of the Macleod sectional map. Obviously, it was not deepened very much at the head. Farther downstream, Verdigris is a striking side-hill river course, the country to the northeast (left) being lower than the channel floor for miles. It also records the vanished ice dam by cutting across a spur of Milk River Ridge, the local slope descending 200 feet in 4 miles against the gradient of the coulee floor. This initial discharge and erosion dates back to the Middle and Kipp coulee stages.

Ponded water of the main Verdigris stage, here named Lake Macleod, was backed up toward the northwest almost to the Little Bow River valley, north of Lat. 50°. Lake Macleod received a large glacial river from the north, traced in this study as far upstream as Calgary. This river crossed the present courses of Bow River, Highwood River, and Mosquito Creek and determined the present course of the Little Bow for its first 25 miles. A few elongated moraine units shown by the Johnston and Wickenden manuscript map occur where the ice edge necessarily stood to hold this side-hill river to its course. The river's gradient was a little more than 30 feet to the mile from Calgary to Lake Macleod. The present Bow River, flowing down the slope of the plains and therefore almost at right angles to the glacial river course, has a gradient near the crossing almost twice as great.

Silt as much as 20 feet thick overlies moraine east of Macleod, making unusually subdued slopes with almost no undrained depressions. Above 3200 feet, however, there is no silt, and morainic topography is normal. Perhaps some of this subdued expression is due to settling and spreading of clay till hills when submerged in Lake Macleod.

Further retreat down the long northern slope of Milk River Ridge next exposed the Etzikom Coulee discharge route, whereupon a drop of about 150 feet in level of ponded water occurred. Verdigris Coulee went dry, all of Lake Macleod was drained, the new lake lying wholly to the east and the head of the new dischargeway being cut in the old lake floor. Present-day Lake Stirling, lying in the coulee head, is at an altitude of 3033 feet. Johnston and Wickenden's small Lake Carmangay is dated from this stage, its outlet channel leading diagonally away from the ice edge and traversing at least half the length of the abandoned floor of Lake Macleod to reach the head of Etzikom. It thus bypassed the newly formed Lake Lethbridge, the main body of ponded water discharging into Etzikom.

Etzikom Coulee, now essentially streamless like Verdigris, is nearly 80 miles long and has a gradient of scarcely 2 feet to the mile. Initial gradients were somewhat greater, for the coulee is 150 feet deep in some places near the head.

Etzikom parallels the southernmost marked moraine belt of Johnston and Wickenden's manuscript map, a belt that laps up to more than 4000 feet above sea level on the north slope of the Cypress Hills in the southeast corner of Alberta. The coulee, 1000 feet lower, discharged into glacial Lake Pakowki at the west end of the Hills. Lake Pakowki in turn discharged southward into Milk River, the discharge crossing the 2900-foot contour in Pendant d'Orielle Coulee and entering Milk River Valley almost at the present river level.

The influence of the Cypress Hills nunatak on the shape of the ice front is obvious both by this southeast divergence of the long marginal river away from the ice edge and by the northeast trend of the moraine units from Lake Pakowki eastward.

With the next change in glacial drainage routes, the Chin Coulee stage, came a cessation of discharge to Milk River Valley and the opening of a route northeastward around Cypress Hills and eastward into Saskatchewan. Chin Coulee is almost as long as Etzikom, nearly parallel with it, from 4 to 10 miles distant and 250 to 300 feet lower. Like Etzikom, it drained from one glacial lake (Taber) to another (Sevenpersons), had an equally low gradient, and is now streamless.

Lake Taber, at an elevation of 2800 feet, lay east of the site of Lake Lethbridge on lower ground exposed by the new retreat. It is the sixth in the succession of glacial lakes, so far noted, whose floors are today traversed by Oldman River; it was 1100 feet lower than Lake Cardston.

The Lake Taber stage includes the functioning of an affluent from the north, this affluent consisting of a chain of three glacial river courses and three glacial lakes which headed at least as far north as Lat. 52° 30′ and descended from 3050 feet to 2800. It may be called the McGregor dischargeway. More than half of its 200 miles consisted of ponded water, the descent being concentrated in the three river courses and averaging a little more than 3 feet per mile. In contrast, the four larger streams crossing this lake-river chain today, and flowing down the slope of the Plains, have gradients from 7 to 10 feet per mile.

The ice edge at this stage was poorly recorded by moraine along the north side of Chin Coulee but is marked along much of the northern lake-river affluent by one of Alberta's most strongly developed moraines, the Buffalo Lake moraine. It should perhaps be termed a moraine system for its maximum width is 30 miles and its ground plan and topography indicate it to be a fusion of several successive linear units. Not until the easternmost unit (Hussar to Hanna) was abandoned and the Forty Mile Coulee route at the south—succeeding Chin and about 200 feet lower at the intake—was opened, did this lake-river

drainage line along the strike of the Plains cease to function. This conclusion is almost certainly correct because, immediately upon the abandonment of the Hussar-Hanna unit, the largest of the three lakes (Drumheller) was drained eastward across the moraine through the premoraine Red Deer Canyon at Dorothy, and this discharge was promptly detoured by the near-by ice front southward along Crawling Valley, the altitude of whose floor indicates it to belong to the Forty Mile Coulee stage.

How far north of Lat. 52° 30′ the northernmost pond of this lake-river extended cannot be determined at present. If the Duffield moraine is possibly a part of the Buffalo Lake system, separated by a big gap between 113° 30′ and 114° Long., then the northern limit of ponding was approximately as shown on the map. However if, as seems more probable, the ice edge extended northward along 113° past Edmonton, ponding must have reached much farther north.[4]

Escape of marginal drainage in part through or under or over a portion of the ice dam does not seem to be indicated or required by field evidence anywhere in Alberta or Saskatchewan. A partial detour of this sort, however, might partially explain the apparently inadequate channel development at Tudor, leading out of Lake Drumheller. As mapped (Rosebud sectional map) at the col, it appears to be less than 100 feet deep, perhaps is but little more than 50, and is only a quarter of a mile wide—disproportionately small for the spillways contributing to Lake Drumheller south of Innisfail. Some water may have escaped from Lake Drumheller late in the stage, under the ice past Dorothy and along the Crawling Valley route to re-enter the surface dischargeway by flowing westward near Bassano. A head of not more than 100 feet was available for a subglacial route totalling at least 30 miles, making the suggestion seem improbable. Heads of as much, and even five times as much, with distances considerably less, existed during the succession from Lake Cardston to Lake Taber, yet there is no suggestion of such procedure in that region.

A third alternative is to continue the use of the Innisfail spillways after the premoraine gorge at Dorothy was opened. It is an unlikely sequence because: (1) The spillways would then be lengthened down Lone Pine Creek valley as far as the 2650-foot contour, and (2) the Burbank Gorge of the Red Deer River would open and drain the Innisfail ponding as soon as ice was off the western half of the Buffalo Lake moraine.

[4] Future students of Albertan glacial drainage derangements will investigate a possible discharge to the Arctic, at the Duffield maximum, of the headwaters of all rivers north of and including the Red Deer.

A fourth alternative assumes that the Hussar-Hanna moraine unit is early Buffalo Lake, the Millicent being the latest ridge of that system. Thus the crossing at Dorothy would occur early in Buffalo Lake time and the Tudor channel would function for only a relatively brief time. This alternative faces the same objection regarding a necessary lengthening of the Innisfail lower channel and also requires opening of the post-Chin Forty Mile Coulee much earlier than do the other views. The essence of the problem is the contrast in Tudor and Innisfail channel development. If the contour mapping at Tudor is poor, a better channel than shown may exist and the problem then would disappear.

Chin Coulee, carrying all this discharge south from Lat. 52° 30′ to about 49° 45′, entered Sevenpersons Creek valley at the west base of the Cypress Hills, thence flowed north and northeast toward Medicine Hat. A curious valleyside channel at 2550 feet lies along the east side of Seven Persons Coulee, connecting at the north end with a marked eastward glacial drainage route now occupied by the west-flowing lower length of Ross Creek. The col in this valley, 10 miles from the Saskatchewan-Alberta line, is at 2500 feet and must be ascribed to known later discharges through the Ross Creek channel. When Chin Coulee water was using it, the floor may have been somewhat higher, perhaps as high as 2550 or 2600 feet.

At Walsh, another ponding was entered, part of Johnston and Wickenden's Great Sand Hills Lake which they show as discharging southwestward to Frenchman River at East End, the eastern end of the Cypress Hills upland. However, Chin Coulee water never used this dischargeway, the col in which is at 2950 feet. It could not have used "The Gap" across the eastern end of the Cypress Hills, a glacial spillway noted by Williams (1929) at 3746 feet, nor the next two spillways to the east of East End (not shown by Johnston and Wickenden)—one occupied by Lac Pelletier, south of Swift Current, whose col is at 2750 feet and the other marked by a short side-hill channel 6 or 8 miles west of Neidpath with a floor at 2650 feet. All these Saskatchewan channels with southward discharge took care of local glacial drainage in Saskatchewan during Etzikom and perhaps Verdigris time.

Not until ice had retreated eastward as far as the Neidpath channel, 23 miles east of Swift Current, could Albertan glacial drainage escape across Saskatchewan territory. Neidpath therefore was the continuation of Chin Coulee discharge and carried in addition all Saskatchewan melt-waters released from the ice front along more than 2 degrees of longitude.

Two tributary glacial river channels enter the Neidpath route, the earlier and presumably Chin-aged one leading eastward from near

Swift Current, and a later one leading southward from the South Saskatchewan River across a large Wickenden moraine unit. Ice stood on this unit when Chin water first flowed to Neidpath. At that time the ice stood also on moraine units a short distance northeast of Medicine Hat, on the moraine between Forty Mile and Sevenpersons Coulees, on the moraine 10 miles northwest of Taber, and on the great continuous moraine extending from the latitude of Bassano to that of Edmonton. This, the Buffalo Lake moraine in central Alberta, is the same age as the Rush Lake moraine east of Swift Current, in Saskatchewan. Thus the position of an ice front, convex toward the southwest, becomes established throughout a length of more than 500 miles.

These five glacial spillways of southwestern Saskatchewan, from "The Gap" to Neidpath, all cross the continental divide between Arctic and Gulf drainage. The divide is most marked in Cypress Hills, becoming lower eastward. This lowering in the direction from which the ice came determined the descending series of successive spillways. There was little northward retreat of the ice front from this divide during their use but more than 100 miles of eastward retreat in the Alberta convexity. The divide determined the existence and location of the southern edge of the Alberta convexity or lobe.

Moraine outlines and courses of two of these spillways farther southward suggest that another ice-front convexity or lobe existed contemporaneously in southern Saskatchewan and northeastern Montana. If so, it was much smaller than the one above noted, hardly 75 miles across from the Arctic-Gulf divide on its north side to the probably unglaciated Wood Mountain (Wickenden, 1931) on its southern margin. The influence of the edge of this lobe in determining drainage routes is detectable only in the short-lived 2650-foot spillway west of Neidpath. It was a side-hill channel, requiring a glacial front on the east for its functioning. This smaller lobe may have been the shrunken remnant of Alden's Great Falls lobe in Montana.

In the region of most rapid retreat in southern Alberta there had been at least six successive lowerings of ponded water by this time, the first five draining across the Continental Divide in Alberta, the sixth finding southward escape in Saskatchewan, more than 150 miles farther east.

This Chin-Neidpath drainage entered another ponded tract, a shrunken remnant of Johnston and Wickenden's Lake Gravelbourg, which now was discharging either through the 2200-foot Willowbunch channel to the Big Muddy or through some more eastern route since obliterated by the building of the Altamont moraine.

Functioning of the lateral channel along the east side of Sevenpersons Coulee, (floor, 2550 feet) southwest of Medicine Hat, required an ice

front along its western margin. This position could not have been held for long; it records the earliest operation of the Chin Coulee stage. Shortly, Sevenpersons Coulee (a fairly broad preglacial valley, not a narrow channel-like coulee), was ice-free and, being lower than 2550 feet, contained a small glacial lake that received Chin Coulee water and discharged eastward along the Ross Creek channel, probably on the same floor used during the immediately preceding lateral channel's functioning.

So long as ice remained on the moraines immediately northeast of Medicine Hat, the Ross Creek channel would presumably continue to function, even though considerable retreat might occur in the convexity to the west. This appears to be what actually happened for a new channel, Forty Mile Coulee, now brought water from the north into the lower length of Chin Coulee, some 10 miles above where Sevenpersons Creek now enters. Though it entered Chin at grade, its three successively used heads near the junction of Oldman and Bow rivers were at only 2700 and 2600 feet—100 to 200 feet lower than the head of Chin far to the west. Thus Lake Taber must have vanished with the opening of Forty Mile Coulee, a new, short-lived lake at 2700 feet and a larger, longer-lived lake at 2600 feet developed farther east, draining southward into Forty Mile Coulee.

Evidence has already been cited to show that this lake (Lake Bassano of Johnston and Wickenden but far more extensive than their map shows) early reached as far north as the Red Deer crossing of the Buffalo Lake moraine at Dorothy. Only an ice barrier could have determined the location of the Crawling Valley glacial stream, and only the 2650- and 2600-foot heads of Forty Mile could have determined the 2650-foot floor of this valley 75 miles distant.

Lake Bassano was long but very narrow in its early stages. It widened as the ice withdrew eastward at least as far as the moraine at Millicent. A few miles of northward retreat west of Medicine Hat then opened the channel now followed by the South Saskatchewan River above Medicine Hat; Forty Mile Coulee was abandoned as ponded water fell another 100 feet to a new level. Apparently this same level was represented by another ponding north of Medicine Hat, one that expanded northward as the ice edge withdrew still farther. Also a new dischargeway became opened, north of and parallel to the Ross Creek channel and at no lower altitude. Lake Medicine Hat included all water standing at 2500 feet in the two nearly separated bodies east and west of Suffield. It overlapped a part of the area covered by Lake Bassano and extended east to within less than half a degree of the Saskatchewan line. Both the Ross outlet and the newly opened Chappice Lake outlet

led to the region of Johnston and Wickenden's Great Sand Hills Lake in Saskatchewan.

The earliest stage of this lake stood at approximately 2500 feet, the altitude of the broad divide into which its outlet is trenched at Neidpath and of the cols in the Ross and Chappice spillways. With the 150 feet of trenching that occurred at Neidpath, the level of the Great Sand Hills Lake was lowered, and the shallow western end drained, ponded water withdrawing east of the Alberta-Saskatchewan line. Because of a lack of contour maps, the outlines of this lake cannot be drawn as completely as have been the earlier pondings in Alberta. Nevertheless, from what the Maple Creek sectional map shows of this lake bottom, it appears at 2400 feet to have been only a series of shallow lakes and connecting channels. No through river ever developed from west to east across this bottom, hence some new and lower route for contributed Albertan run-off must have been open when the Neidpath discharge ceased.

Thus an "Early" Wisconsin moraine (Chin Coulee-Buffalo Lake stage) has been traced around the Albertan convexity and eastward to Long. 107°, its course in the last 100 miles being somewhat north of east but the eastermost unit (south of Rush Lake) curving around to point far to the south of east. The next moraine east of the Rush Lake unit is the great Coteau or Altamont moraine, of late Wisconsin age, its course aligned nearly northwest-southeast for more than 100 miles in both directions, thus departing some 50° from the course of near-by Chin Coulee moraines recording the ice front at the Neidpath stage. This angular discrepancy and also the disappearance of the westward convexity record a notable change in outline of the Keewatin ice when it readvanced to build the Late Wisconsin moraine. Despite the southeastward curving of the Rush Lake unit, the ice edge of the Chin Coulee stage probably continued eastward from Rush Lake, its record now lost beneath the younger Altamont.

Sags among the Altamont units are consistently well below the Neidpath channel, yet none carries record of east-flowing glacial water. Since they do indicate the eastward descent of the plains, probably some now obliterated escapeway lower than Neidpath appeared as the ice front withdrew from the Chin Coulee moraines, Neidpath being abandoned in its favor.

During the late functioning of the Ross and Chappice spillways, therefore during the life of Lake Medicine Hat and Great Sand Hills Lake, the ice front shrank back to the Oyen moraine, leaving the great moraine belt crossed by the Red Deer River between Steveville and Empress. About midway of this retreat, specifically at the Buffalo-Cavendish-

Pivot moraine stage, a small lake may have formed between the Bowmanton moraine north of Medicine Hat and the ice edge. The only evidence for this is the Red Deer Lake valley which, though undoubtedly premoraine in origin, has an evenly graded floor at about 2450 feet, as though it had been used by a large glacial stream during ice retreat. Any discharge through it was southward to the west end of Great Sand Hills Lake. A difficulty this hypothesis encounters is that the present South Saskatchewan River course, north from Medicine Hat, must be determined by a premoraine valley, lower (despite moraine deposits) than the Ross and Chappice channels and presumably as low as the Red Deer Lake valley. Perhaps it was not quite that low. Perhaps the lake had two simultaneous discharge channels, like Lake Medicine Hat.

By the time the Empress region about the Red Deer and South Saskatchewan junction was exposed, Medicine Hat and Great Sand Hills lakes had been drained, and the Neidpath channel was dry. The South Saskatchewan then ran across the abandoned Medicine Hat Lake bed, flowing north from Medicine Hat across the Bowmanton and the Pivot-Cavendish moraines by way of a premoraine valley lower than Ross or Chappice channels. It entered a glacial lake covering the recently exposed Empress region. Silts about Empress overlie ground moraine up to at least 200 feet above the river, perhaps 300 feet, and glacial ice may have been on the Alsask moraine at the time. The early lake may have stood as high as 2300 feet. This, however, would not put it in the Neidpath group of pondings, and therefore the ice margin was well back from the Rush Lake moraine unit, at least as far north as the South Saskatchewan-Swift Current junction. A dischargeway, now obliterated, was crossing the Missouri Coteau somewhere near where the South Saskatchewan now crosses. East of the Alsask moraine unit, Wickenden's manuscript map shows no morainic record of the position of this portion of the Lake Empress stage's ice front. Because of the lack of contour maps for a large part of the area covered by Lake Empress, its outline cannot be shown. Johnston and Wickenden have a lake outline on their map, but they have clearly included all standing water stages from the Lac Pelletier (2717 feet) discharge to Lake Empress, and their outline cannot be used on the present map.

This correlation of moraine belts of Alberta and Saskatchewan, based on the sequential operation of their melt-water rivers, necessitates acceptance of varying rates of recession from the Alberta convexity, especially marked from Lake Taber to Lake Empress time. On an axial line drawn through Empress and Lethbridge, the ice retreated northeast for 150 miles, while at the east end of the southern flank of the convexity it withdrew toward the north scarcely more than 25 miles—only

one-sixth as far. If this is correct, the near-by moraine units east and north of Swift Current contain, in a width of 10 to 12 miles, the age equivalents of every moraine in Alberta from the Buffalo Lake to the Sibbald, a total of 10 such belts.

In contrast are the indicated retreatal rates during pre-Chin, pre-Lake Taber time. Correlations indicate a withdrawal along the Continental Divide, which determined the southern margin of the convexity, of 100 miles, while in the lobe front it was retreating only about 60 miles. This greater recession in the re-entrant was in essence the emergence of the higher parts of the preglacial divide through the thinning ice. After such emergence the southern margin of the lobe consistently shifted down the divide's northern slope to lower ground. Here its retreat might well have been much slower, and its replenishment with Keewatin ice much more rapid than either had been on the divide upland. In these possibilities may lie part of the explanation for this reversal of retreatal rates in the convexity and the re-entrant.

In contrast with the abundance and unequivocal glacial river origin of the channels in southern Alberta and Saskatchewan is their paucity and dubious character north of the latitude of Empress (51°). A few large and continuous units to the north take the place of the many scattered discontinuous moraines mapped by Alden, Johnston and Wickenden, and the writer south of that line. There is therefore little need of glacial drainage routes to establish correlations in the north. The only outstanding correlation problem here concerns the farther northward course of the ice edge recorded in the Millicent-Sibbald series of moraines. Possibly the Millicent is an isolated part of the back edge of the Buffalo Lake moraine system and should be tied to the moraines south of Lake Newell and among the heads of Forty Mile Coulee instead of to the Agatha moraine southwest of Suffield. This would necessitate only slight changes in the sequence of lakes and outlets already outlined. Possibly the Millicent and the Jenner-Buffalo-Cavendish-Chinook-Oyen units have no existing correlatives farther north, though faint ridgings were seen in the field in positions which might mark such correlatives. Also, all but the Millicent may belong to the big Veteran moraine.

Williams and Dyer (1930) thought the Kerrobert glacial river channel, 60 miles long in western Saskatchewan extending approximately from Lat. 51° to 52°, drained northward to a glacial lake. This channel appears on the Kindersley sectional map which has a 100-foot contour interval. A 50-foot interval would bring it out strikingly. Williams and Dyer say it is 150 feet deep. Part of its floor is below 2,000 feet, part above, but the map does not make clear the direction of flow.

Traced northward onto the Tramping Lake sectional map (50-foot interval) a channel-like continuation can be traced as far as Macklin, its floor below 2300 feet throughout, much of it below 2250 feet. From Macklin it can be traced into a complex or irregular chain of lake and swamp basins on the Sullivan Lake and Wainwright sectional maps but without any one obvious channel route through the maze. North of this the country is much smoother and rises gradually to something above 2250 feet in the lowest place on the divide between Battle and Vermilion rivers. The 2300-foot contours cross the divide, about 10 miles apart. However, there is no suggestion of a channel form here nor down the smooth gentle northern slope of this divide into the Vermilion drainage, descending 150 feet in 25 miles. Therefore the Kerrobert channel must have drained southward, probably from a very irregular body of standing water in front of the Altamont moraine between Macklin and Viking. It may have originated as an "Early" Wisconsin watercourse but almost surely was used during Altamont time. Its southward-draining water did not, at either time, go through the Neidpath channel (2350 feet). Nor did it escape across Log Valley pass to Chaplin Lake (2170 feet) for the lowest land north of Chaplin Lake is 2350 feet, and no channel crosses it.

Only one route appears possible: The Aquadell channel on the Rush Lake sectional map, lying east of the main Altamont moraine and south of the South Saskatchewan River. It is about 150 feet deep, is cut longitudinally along the eastward slope of the Coteau, and has a floor at an altitude just under 2000 feet. Its discharge was southward. The South Saskatchewan valley could have carried water from the south end of the Kerrobert channel to the north end of the one at Aquadell, descending nearly 200 feet in the 100-mile interval.

This is a permissible hypothesis providing that the 550-foot canyon of the Saskatchewan here is postglacial. Otherwise, a narrow lake would occupy three fourths of this distance and would increase the gradient in the remaining one fourth to a figure incompatible with the map evidence.

This functioning of Aquadell must have succeeded that of Chaplin-Johnstone-Lake of the Rivers discharge on the west side of the main or inner Altamont and must have coincided with the last stand of retreating ice on the inner Altamont's back slope, south of the river. The position and altitude of the south-draining Whitebear channel, 25 miles north of the Saskatchewan gorge, strongly suggest that the edge of the retreating Altamont ice was not aligned as is the Altamont moraine, that it first withdrew south of the Saskatchewan crossing. Thus the Aquadell spill began while Whitebear was still buried under the glacial margin.

The puzzle of where drainage went at the Altamont maximum is still unsolved. No adequate southward escape route other than through Neidpath is known, and this seems too small for both "Early" and Late Wisconsin discharges. A further puzzle is the source of the large quantity of water that used the Chaplin-Johnstone-Lake of the Rivers route. Alden (1932) speaks of the Big Muddy discharge in Montana (to which this led) as a main through valley for glacial water. The map shows that only about 75 miles of ice front at the second Altamont stage could have contributed directly to the river that went out through Lake of the Rivers channel.

Possibly a broad shallow spill southward across Log Valley pass and into Chaplin depression began from the Saskatchewan valley as soon as ice had retreated to the inner Altamont. This hypothetical spill left no channel, the absence of which may be explained as follows: Lake of the Rivers is the farthest downstream of the three lakes and is the only one lying in a real channel. Its altitude is 2203 feet, its clean-cut bluffs are 100 feet high. The country into which it has been cut is a rather smooth plain, 2300 feet and higher near the north end of the Lake of the Rivers channel. Thus both Chaplin and Johnstone depressions contained a glacial lake at 2300 feet at the initiation of this hypothetical discharge. The Log Valley route then had a descent of less than 50 feet over a distance of about 15 miles. Perhaps no channeling would have occurred in Log Valley under these conditions, and possibly the Lake of the Rivers deep valley was cut after the Aquadell channel opened and Log Valley went dry.

Another hypothesis explaining the volume of water down Lake of the Rivers channel is that the "outer" Altamont west of this channel is really Early Wisconsin moraine. If so the "inner" Altamont unit between Chaplin and Aquadell is the same as the "outer" unit north of the Saskatchewan. The retreating ice front would then allow opening of the Aquadell channel while Whitebear was still ice-covered. Neidpath would not necessarily carry any Late Wisconsin water if we can accept the 2350-foot tract at Log Valley as a spillway.

Thus the Kerrobert channel was a Late Wisconsin dischargeway, as must be therefore all Saskatchewan channels farther east. They all may have had an earlier history, but, if so, it cannot be deciphered today.

PRE-WISCONSIN DRAINAGE IN ALBERTA

Aside from the glacial drift and the trenchlike river valleys, southern and central Alberta is a fairly smooth plain across the eastern limb of the Albertan syncline, underlain by late Cretaceous and early Tertiary sediments. The slope of the plain is eastward and northeastward, the

dip of the underlying beds is westward. Several thousands of feet of sedimentary rocks are beveled by this plain which thus is clearly erosional in origin. Residual hills and groups of hills are scattered across the plain, most of them less than 500 feet high, some 1000 feet, and Cypress Hills more than 2000. The cap of Rocky Mountain gravel on Cypress and Hand hills records early stream erosion across the edges of the warped rocks of the plains. Continued erosion has lowered the surface until the existing oldland has been produced.

Except for the laccolithic Sweetgrass Hills on the 49th Parallel, and perhaps the gravel-capped Cypress and Hand hills, none of these residual elevations contain rock more resistant than what was eroded to leave the eminence. These monadnocklike residuals are but surviving interfluve tracts of the older drainage, and the farther from them the main rivers of their cycle can be placed, the more satisfactory the reconstruction will be.

However, the drainage pattern today is not that of the late Tertiary oldland.[5] Many of today's main rivers cross some high tracts of this oldland, following routes that are obviously aberrant for the older drainage. Williams states for one case that ". . . the land in the proximity of the river . . . is higher than the land at some distance from it." Many instances of this show clearly on the contour maps, many can be recognized in long-range views in the field. Some of these aberrant courses are moraine and ice-front deflections dating from the Wisconsin glaciation. Numerous other such valleys, though definitely postoldland, contain morainic accumulations or are so situated and so oriented that the retreating Wisconsin ice front could not have determined them. It seems that such valley courses must have been determined during earlier glaciations, almost all other records of which were obliterated during Wisconsin time.

The Red Deer River course on the plain may be taken as an example. In two deep gorges it crosses one of the largest residual uplands, here called the Delburne upland. In another such gorge it flows between two closely set uplands (Wintering Hills and Hand Hills) whose summits are respectively 1150 and 1350 feet above the river, though only 5 and 12 miles distant. Conclusions in the following simplified presentation seem to the writer to be amply supported by facts which, in the interest of brevity, are omitted.

The northeastward-flowing Red Deer River, above the town of Red Deer, is in its preglacial course and is little, if at all, entrenched into

[5] Williams and others have noted the occurrence of some drift-filled valleys, but data of this kind are too few for any satisfactory reconstruction of an earlier drainage system. In this paper another kind of glacial derangement of the older drainage system will be briefly considered.

the oldland. Downstream a few miles from this town, the river leaves its old course on the broad lowlands, turns abruptly eastward, traverses a gorge 600 feet deep and less than a mile wide at the top across a preglacial divide of this upland, and enters the headwaters of a minor preglacial valley down the east side of the divide. The gorge is incised in one of the highest parts of the upland, drains in the direction from which the ice came, could not have been initiated during the building of or retreat from near-by Wisconsin moraines. Broad unchanneled places occur in this divide 200 feet lower than that where the river crosses, yet only a few miles distant. Moraines or ice margins of earlier glaciations, with some other outlines than those of the Wisconsin stage, seem necessary to initiate the crossing.

Where almost across to the lowland on the east, the Red Deer turns southward, re-enters the upland in a gorge whose depth increases, because of increasingly higher land, 300 feet in a distance of 8 miles. In this distance the stream by its descent adds only 50 feet to the depth of the gorge. This southward to southeastward course of the second gorge parallels the courses of half a dozen maturely proportioned minor preglacial valleys down the eastern side of the Delburne upland. It seems clear that one of a group of small valleys (the Ghostpine group) here was entered by way of another divide crossing and has been made over into a major discharge route across the plains from the mountains. Rejuvenation, proper to this view, has occurred in the lower stretches of the otherwise only slightly altered surviving members of the group.

Remaining in a trenchlike valley but with a wider floor, the Red Deer continues for 30 miles farther to reach the Wintering and Hand hills where, although the land immediately back from the bluff summits does not rise downstream, the two hills are less than 20 miles apart, one on each side of the river trench. Here a moraine crossing is well shown, dating the trench as pre-Wisconsin. Here also is a broad erosional terrace on either side, about half way down from the Hills summits to the river level. The surface of this terrace, or berm, rises gradually upstream into the preglacial valley floors of the Ghostpine stream group on the Delburne upland. Downstream, it is continuous with the broad gentle oldland slopes of the Bullpound Creek drainage.

Here seems to be a slightly modified unit of the oldland's drainage system. Wisconsin moraine deposits have been responsible for the Sullivan Lake ponding at the head of Bullpond. Pre-Wisconsin ice (and perhaps moraine) seems to have detoured the Red Deer across two divides of the upland and southward into the Ghostpine group to make the gorge and to cause the rejuvenation all their lower courses show. The Red Deer's own rejuvenation has extended upstream no farther than

the town of the same name. Since all the larger rivers of this part of the Canadian Great Plains are in youthful valleys much like the Red Deer, pre-Wisconsin tilting of the plains has probably steepened the eastward slope.

Insufficient data are at hand to attempt further unraveling of Alberta's pre-Wisconsin drainage changes. Battle River offers a problem in that it apparently has lost, rather than gained, volume since the rejuvenation. It receives no Rocky Mountain drainage today, yet quartzite gravel occurs beneath the drift along its valley slopes. Battle River may have carried Red Deer drainage before the Delburne upland's gorge routes were initiated, may once have received the North Saskatchewan discharge via Coal Lake Valley. Vermilion River valley downstream from Two Hills appears to have been the preglacial course of the North Saskatchewan which is now entrenched in an upland somewhat farther north. There are many more such anomalous stream situations for which the Wisconsin glaciation cannot have been responsible.

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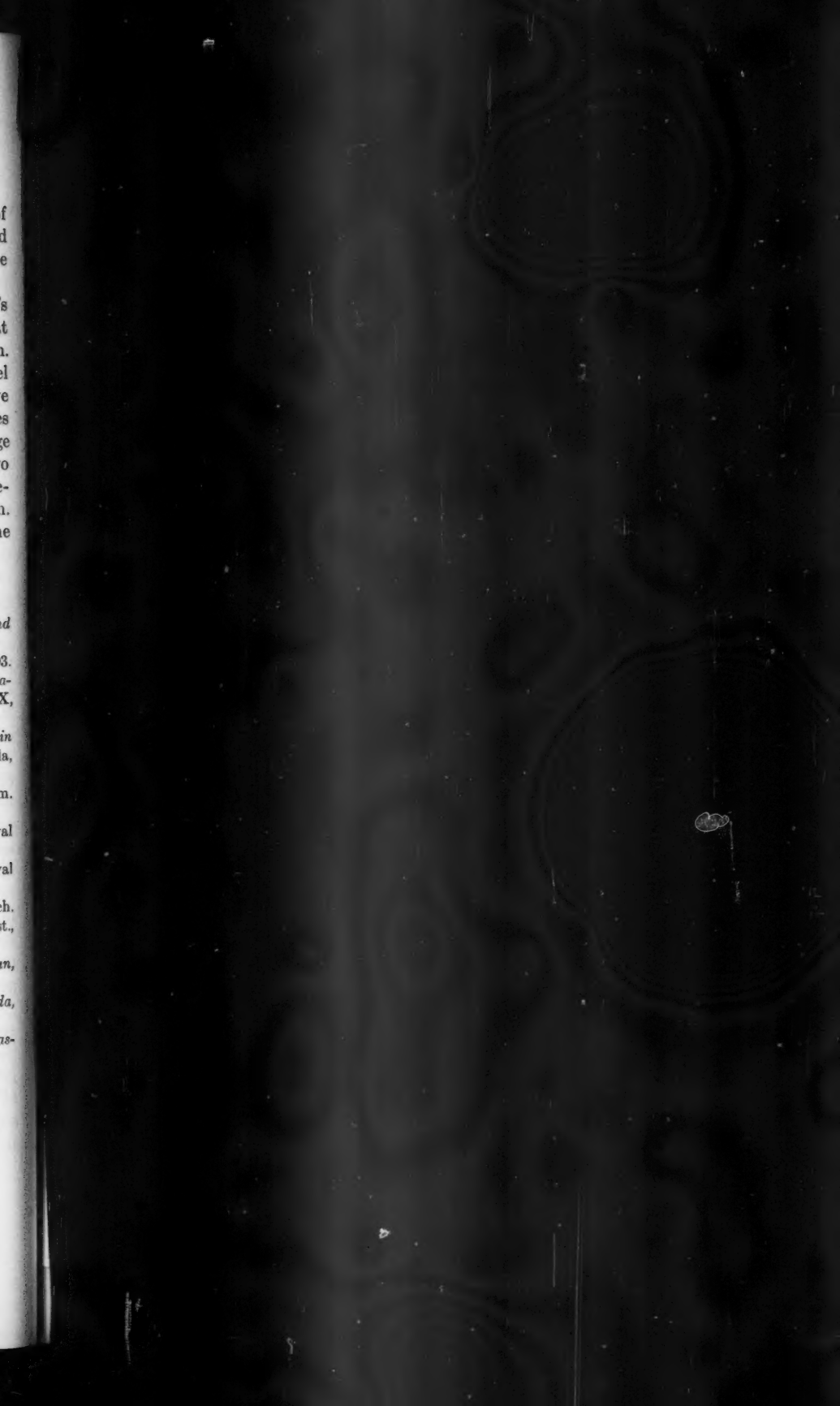

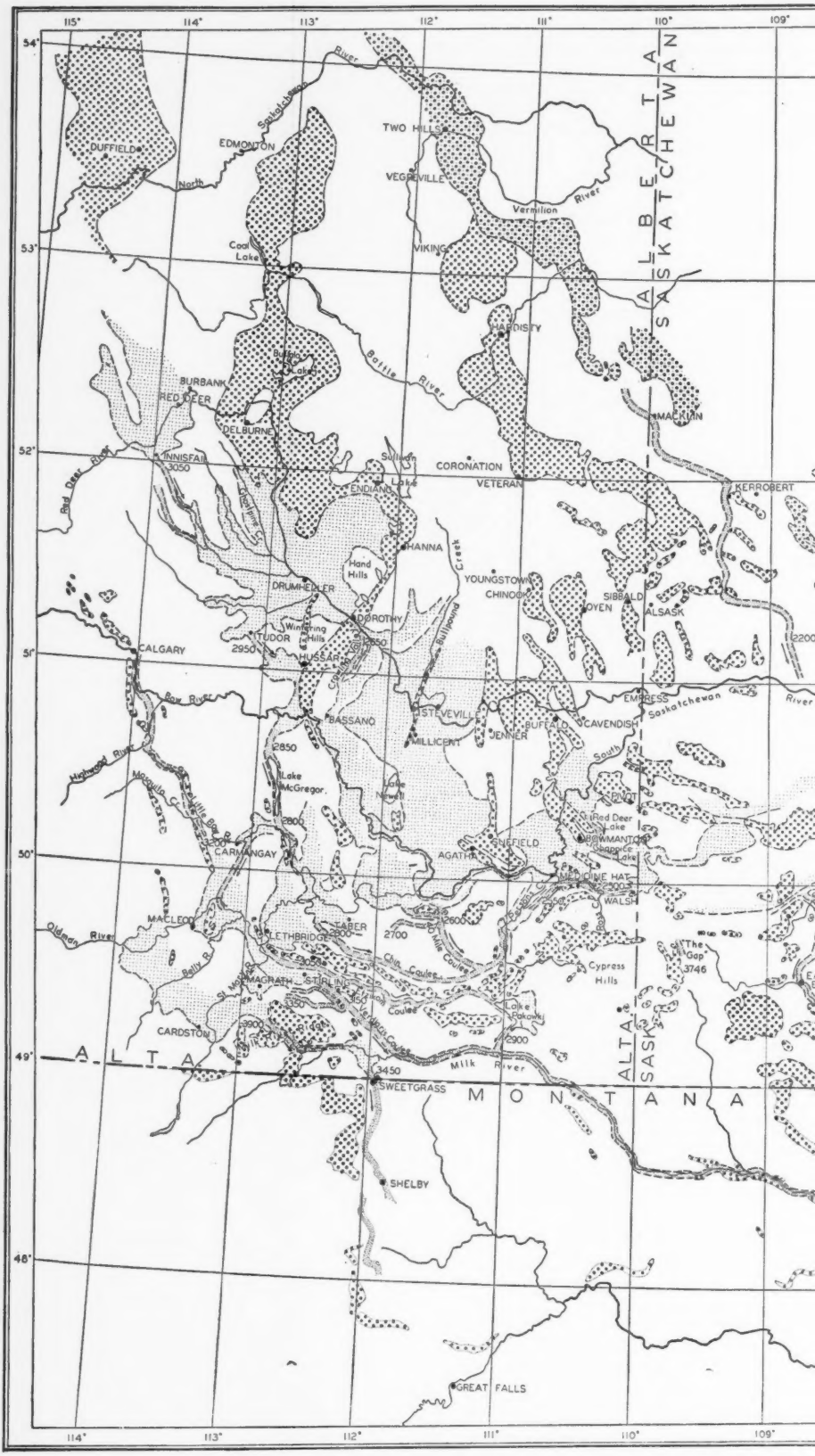

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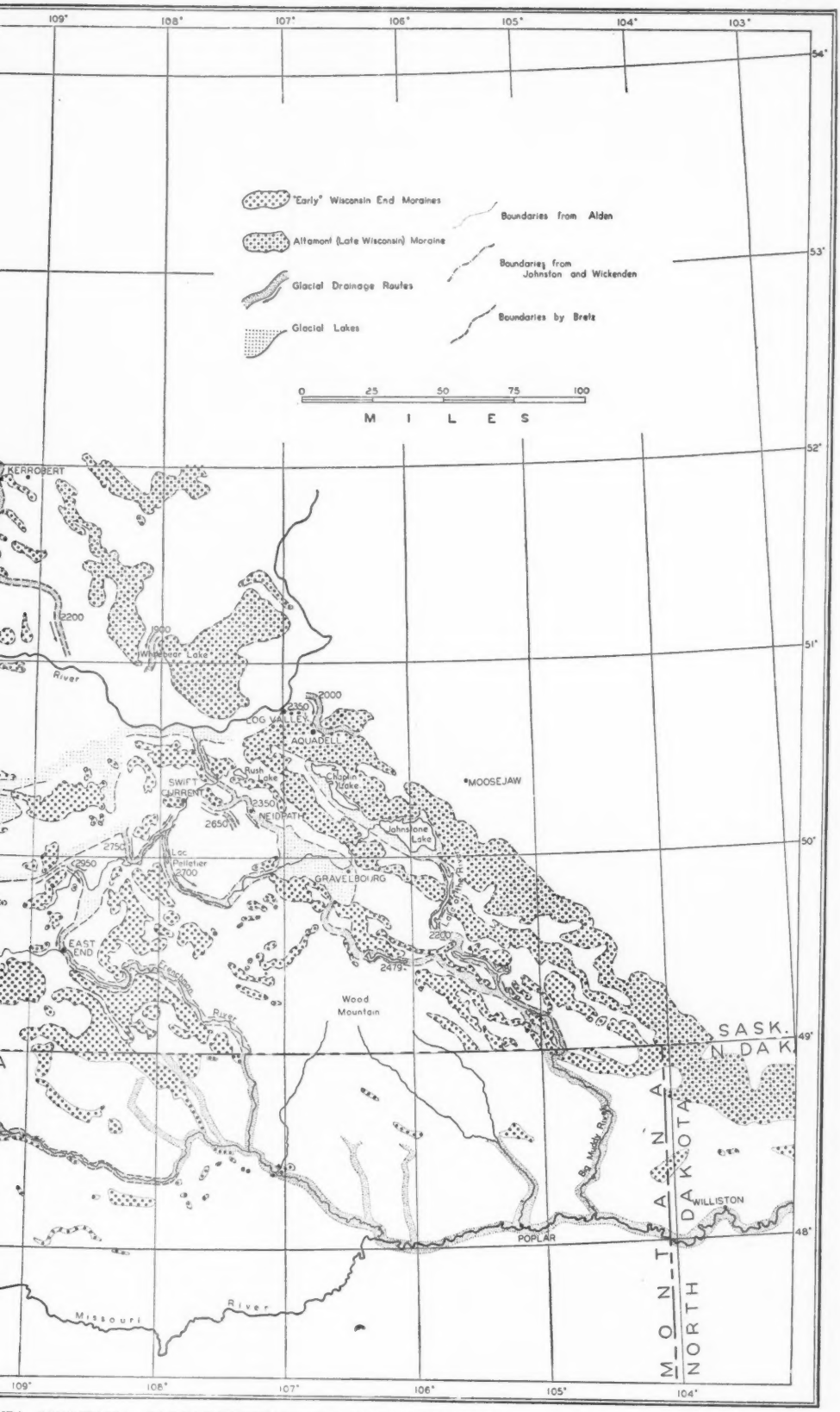

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SOIL AVALANCHES ON OAHU, HAWAII

BY CHESTER K. WENTWORTH

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ABSTRACT

A large part of Oahu above 1000 feet in elevation is very rugged, with steep slopes from 35 to 60 degrees. In the rainier parts, though widely distributed, bare rock cliffs are usually not of great lateral extent and occupy but a small part of the map area. Nearly the whole area is covered with soil or rock so weathered as to support a dense cover of vegetation. By incessant creep and colluvial wash, this mantle rock assumes slopes at the limit of stability for normal conditions. In exceptionally wet weather, owing to weight of soil water and the lubrication it affords, patches of soil together with vegetation become loosened and pass downslope as soil avalanches. The angles of slope are chiefly from 40 to 50 degrees, and the thickness of material involved is rarely more than 2 or 3 feet.

The avalanches play an important part in the succession of plants, but the details of this phase have not been studied. They are also thought to be important in undercutting the crests of ridges so as to develop the striking knife-edge ridges so

characteristic. No close measurement of the rate of erosion by these slides is practicable, but they apparently are responsible for a relatively rapid denudation and they are probably the controlling factor in determining the general form of land sculpture in the mountainous parts of this region.

INTRODUCTION

SCOPE OF OBSERVATIONS

This description of soil avalanches is based on nearly 8 years of field study of the rugged topography of the Koolau Range, which in part forms the Honolulu Watershed, and considerable knowledge of other parts of Hawaii. Moreover, it deals chiefly with contemporary landslides, many of them visited within a few days or hours of their occurrence, and for which the general conditioning factors are well known.

Observations were chiefly made on the valley sides of the leeward slope of the Koolau Range in the District of Honolulu, from Waialaenui Valley on the east to Moanalua Valley on the west, in the course of detailed areal mapping of the geology of the watershed. In carrying on this survey at the scale of 500 feet to the inch, a large number of traverses have been made in the more mountainous country inland from Honolulu, most of them involving the use of a rope in passing down the head or side walls of the valleys and affording abundant opportunity both to identify newly developed slide scars and to visit the sites at various times after the sliding.

Unfortunately, from the standpoint of complete understanding of the slides, the survey work is done when the weather is most favorable, with the least interference from rain and drifting fog. Consequently, despite unavoidable, frequent encounters with heavy rain, neither the writer nor his assistants have actually seen a slide take place. On one or two occasions noises attributed to such slides have come from a mile or more away, and slides not over 1 or 2 days old and rather obviously related to extraordinarily wet weather have been observed commonly.

ACKNOWLEDGMENTS

For general aid in connection with the survey, thanks are due several field assistants, including especially Austin E. Jones and Horace Winchell. Parts of this manuscript were first drafted about 1936, prior to the publication of *Landslides and related phenomena* by Sharpe; additional observations have been stimulated by reading that work.

DESCRIPTION OF THE SLIDES

PHYSIOGRAPHY

The Koolau Range inland from Honolulu rises to crest elevations of 2000 to 3000 feet, and its leeward slope is dissected by a succession of deep valleys, 4 to 6 miles long. The remnant spurs separating the

several valleys, near the coast and inland to elevations ranging from 900 to 1300 feet, consist of sloping, triangular facets, which are little-dissected remnants of the constructional surface of the original lava dome. At an average distance of about 1½ miles from the coast, each of these flow-slope facets narrows to a sharp apex, inland from which the intervalley divide continues as a steep-sided, narrow, "knife-edge" ridge. This is commonly somewhat serrate, with peaks and sags differing in elevation by 100 feet or more, while it rises in an inland direction at approximately 400 feet to the mile. It is clear that the crests of these lateral ridges are cut somewhat below the original surface of the dome, especially in the zone where the valleys on either side are deepest and before one reaches the steeper, head portion of the valley which rises to the crest of the range.

Much of the inland, mountainous portion of Oahu conforms to this general description. None of the valleys is flat-bottomed farther inland than at 400 or 500 feet in elevation, except those that have been lava-filled. Lower valley walls are flanked by massive talus fans, which in places are veneered over bedrock buttresses. Higher valley walls are steep, 45 to 70 degrees, and more or less corrugated by alternate ribs and flutings (Pl. 1, fig.1). Most streams have few tributaries other than those in chute types of channel developed on the steep valley wall. Such tributary valleys as occur are often subequal, consequent valleys which are essentially parallel to the main valley. Some of the major streams at the head branch out into a palmate system of radiating channels which come radially down the moderately dissected, blunt head of the valley.

It is emphasized that a large part of the mountainous area of Oahu, where rainfall exceeds roughly 80 inches and where there is a dense cover of vegetation, consists of slopes typically ranging from 35 to 60 degrees. These are usually somewhat fluted, with the ribs rising 100 to 400 feet above the chutes (Pl. 1, fig. 1). In slightly drier or intermediate rainfall zones there are commonly some outcrops of rock along the axes of the ribs, but in the more typical, wetter areas the larger rock outcrops are chiefly and often disconnectedly in the chute channels.

GEOLOGY

Except for the coastal lowlands of reef limestones, tuff and cinders of secondary cones, and sedimentary derivatives, the entire island of Oahu is formed of thin, basaltic lava flows. Most of the flow units are no more than 10 feet thick, and characteristically they consist of dense interior parts, with scoriaceous, somewhat detrital upper, and in some places lower, portions. Thus, the structure, in sections in drier valley

walls where rock exposures can be seen, has a stratified aspect with an alternation between the bolder ledges of the dense parts and the slightly undermined or overgrown scoriaceous layers which weather more readily and furnish more foothold for plants. The lava flows slope in most districts in a general seaward direction 300 to 700 feet to the mile and parallel to the slope of the interfluve ridge crests.

In the rainier sections where rock outcrops are mostly in the steep stream channels, there is less contrast in exposure between the dense layers and the clinkery layers, but here also the steep and better exposed parts are often occasioned by the thicker and denser lava flows. In this rainy area, to which the present paper especially applies, the proportion of available outcrop is very small, estimated as not over 2 or 3 per cent. Exposures are commonly much weathered. In some of the larger stream channels nearly fresh rock may be found, especially in thick flows or in dikes, but in many places specimens for petrographic study can scarcely be obtained.

Most of the remainder of the surface is covered to varying depths by mantle rock which consists of an ill-assorted mass of detritus ranging from large blocks to silt. Because of its generally weathered condition, this mantle rock, which the writer has designated as "taluvium," is comparatively impervious. Some parts are somewhat porous, and this forms material, together with the underlying weathered bedrock, retards the infiltration of ground water and thus helps to support the dense vegetation. Were they not so covered, the naturally very pervious fresh lava flows would promote a very rapid passage of rain water downward to sea level.

Adjacent to the lower courses of larger streams, this taluvium forms large talus masses whose upper edges rise several hundred feet above the stream channel (Pl. 1, fig. 1). In many valleys these cones reach alternately out across the valley floor from opposite valley sides, causing the stream channel to follow a markedly meandering course around them. In other valleys, and in the narrower, upper courses of most valleys, the taluvial mantle on the valley walls is thinner and is either cut off sharply below by the stream channel or merges with a slight alluvial cover on the narrow valley flat.

VEGETATION

Only brief mention of the vegetation can be offered here. In a recent ecological study of Kipapa Gulch, in which vegetation is similar to that of the Honolulu Watershed, Hosaka (1937) has distinguished six plant zones. From the leeward coast to the valley head, these are named Maritime, Haole Koa, Guava, Koa, Ohia, and Cloud. The area in

which rainfall is sufficient so that soil slips are common lies chiefly in the latter three of these. A detailed description of the vegetation is presented by Hosaka. In passing inland from the coast, the layman notes the gross transition from cactus and lantana on the drier slopes to guava on moister valley sides, koa on intermediate ridges, and then ohia in the higher mountain sections. A striking feature is the marked prevalence of the kukui tree, with its conspicuous light-green foliage, in the axes of valley-side gulches and middle parts of the main valley (Pl. 1, fig. 1). Far inland in many valleys are extensive steep slopes covered by so-called staghorn fern, a plant which appears to be expanding its range at the expense of the forest.

In some valleys, such as Nuuanu, Kalihi, and Moanalua, there are considerable areas of grasses. Along streams, and on flat, moist valley-bottom areas, the hau forms an amazing, interlaced jungle of stems from an inch to a foot in diameter. The various antecedent and subsequent relationships between slides and specific plants would constitute a fruitful study but has not been attempted here.

SIZE AND LOCATION OF SLIDES

The type of slide here described appears to be the common, contemporary sort. At least two other varieties are known. One occurs when a great mass slumps off a high sea cliff, forming the large fans which are seen, and mapped, at the base of certain cliffs on Lanai, Molokai, the Waipio district of Hawaii, and elsewhere. Another is that occurring from the walls of Halemaumau, at Kilauea. Stearns (1935) has mentioned certain "mudflow" deposits which might also be considered in this category, but the present writer believes that some, if not all, the coarse, ill-sorted material ascribed to mudflows by Stearns is akin to the typical taluvium formed by successive, ill-sorted gravity and colluvial additions to the cones which flank the valley walls.

Slides of the current type are located on nearly uniform slopes of 40 to 50 degrees. A compilation of measured angles shows 80 per cent of the total between 42 and 48 degrees, with an occasional slope as high as 55 degrees. The scar left by the removal of vegetation is conspicuous and can readily be seen from the coast to the crest of the range in clear weather. The accumulation of material in the valley or chute channel is inconspicuous from a distance and is chiefly significant as a temporary derangement of slopes and inconvenience to travel. The writer has seen only very small areas of landslide topography due to these thin slips.

These slides range chiefly from 100 feet up and down slope by a quarter as wide, to perhaps five times as large in linear dimensions. Some slide directly into main stream channels from lower valley wall

slopes. Others, and perhaps more, slide from higher positions on the valley wall, their tops reaching a point immediately over the side from the knife-edge ridge. Some of these are located near the center of the chute head; others begin on the lateral slop of a rib and slide into the mid part of the chute (Pl. 1, fig. 2).

It is significant that these slides do not, in general, jeopardize the

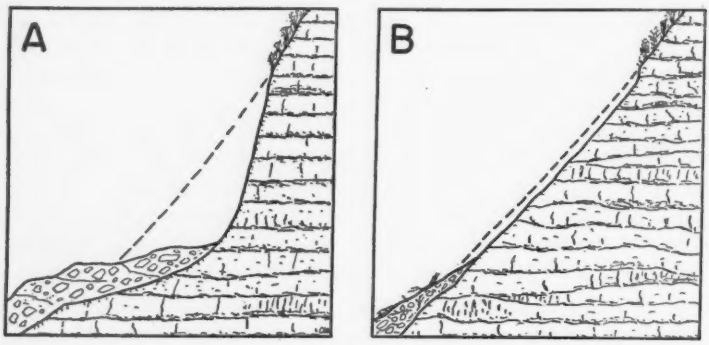

FIGURE 1.—*Diagrams showing difference between (A) slumps, and (B) soil avalanches*

protruding upper edge of the steeper, rock walls. They are essentially soil, or humus slides, which do not involve sound rock nor affect the rock slope except indirectly (Pl. 1, fig. 3).

These soil avalanches are sharply to be distinguished from the type called slumps by Sharpe (1938, p. 66-67), which are due to shearing within the soil or rock mass caused by pressure-cumulated stresses. The soil avalanches are due to frictional failure on a surface close to and essentially parallel to the topographic surface. (*See* Figure 1.)

ANGLE AND THICKNESS

A compilation of measured angles of slide scars shows almost a third recorded as 45 degrees, and 80 per cent between 42 and 48 degrees. It is apparent that this general declivity is one of marginal stability for such soil and mantle rock, with its particular vegetation and roothold. Thus, in the wettest weather, with a decrease in friction due to saturation and an increase in weight of perhaps 500 pounds per cubic yard, the mass is unstable, but during the remainder of the time it is stable.

Thickness of the material dislodged in these slides is commonly 1 to 2 feet, rarely as much as 3 feet, as readily seen along their lateral margins. In general, the slides are neither thicker nor thinner at the top than at

the bottom. At margins, they may slightly encroach on areas of other slopes, but as a rule they appear to bring about a successive baring of the surface, without significant cumulative change in the slope (Pl. 2, fig. 1).

RELATION TO ROCK

Rock in place is commonly observed in the scar exposure left by a slide (Pl. 2, fig. 1). In some scars, rock appears in places along the strike of the slope where a more resistant and less weathered flow ledge occurs, but between such points taluvium or more obscure residuum is exposed. Rock surfaces exposed in slide scars are commonly smoothed and striated (Pl. 2, fig. 2). Except for the slight irregularities in rock surface due to variations in weathering, the rock surface underlying most of the slides is closely parallel to the topographic surface before and after the slide.

TERMINOLOGY AND CAUSE

With reference to the descriptions and classifications compiled by Sharpe (1938, p. 61-63), the slides here discussed are considered as soil avalanches. The writer prefers the word soil to debris as used by Sharpe, because in this rugged topography the mantle rock which slides is the soil on which a lush, dense vegetation grows. The slopes of these slides are prevailingly much higher than those mentioned by Sharpe (20 to 40 degrees), but the clinging of mantle rock to slopes of 40 to 55 degrees is probably due to dense growth and subtropical conditions of weathering.

Repeated observation shows that soil avalanches take place chiefly during periods of heavy and prolonged rainfall. Records of rainfall at a rate of 3 to 5 inches per hour are known for various places in the Hawaiian Islands, and it is believed that local intensities of 10 inches per hour for a quarter or half hour are occasionally reached in the higher mountain section where few records are taken. In particular, areas of 100 or 200 acres have been observed in which several slides have taken place and in which it is evident that excessively high rainfall intensities have obtained locally. It seems clear that both the added weight due to water saturation and lessening of friction by water are the immediate causes of sliding, on slopes which are built up to near the critical slope by continued creep and colluvial movement of soil and rock detritus.

GEOMORPHIC SIGNIFICANCE

FUNCTION IN SLOPE CONTROL

The close parallelism between the topographic surface and the rock surfaces under slides, and the generally close parallelism of the two in much of the upland surface, except in valley bottoms and under some pronounced taluvial fans, indicate that a considerable part of the rock

surface is weathered and eroded to a declivity which induces instability of mantle rock in the wettest weather but on which mantle rock generally lies. Rock surfaces so steep that they remain bare of detrital accumulations are everywhere subject to slow wastage, and unless they are maintained by erosional encroachment at the base are destined to become lowered by accumulation of talus at the base, by erosion at the brow, and by rise of the base through pediment formation at the angle of adjacent talus stability. Such cliffs or bare rock bluffs may be the result of forces no longer active and left as waning remnants, or they may be the local expression of particularly resistant rock in situations where they are more or less continuously maintained. In any event, such surfaces constitute only a small fraction of any map area. All other surfaces, developed on mantle rock, can be no steeper than the angle at which mantle rock is stable. Much of the Koolau Range is cut by deep valleys, and the weathering of the basalt on the upland surfaces is deep and produces much debris; hence much of the valley-wall area should be mantled by rock debris disposed at near the limiting angle of stability.

Mantle rock overlying a bedrock surface does not protect the bedrock from the chemical weathering which extends downward several scores of feet in some places, but it does protect the rock surface in large measure from disintegration and reduces the amount of differential etching out of hard and soft layers which elsewhere develops on more exposed rock surfaces. There is evidence that in various places great fans of talus which flank the bases of steep cliffs are underlain by rock buttresses which seem to have been protected while the cliff face has receded. Once granting that rock faces too steep to retain a detrital mantle recede horizontally more rapidly than the gentler rock surfaces covered by mantle rock, it is apparent that the slopes at the angle of mantle rock stability will gain in area over cliffs. It is also apparent in a region of high rainfall and active runoff that mantle rock is subject to addition from upslope and to loss below, hence that it will tend to assume slopes as steep as will be generally stable, while approaching ephemeral instability. This is essentially the same principle as that which determines the grade of a given stream channel at a given place. The debris-mantled slopes tend to become steep enough to enable mantle rock to move downslope as freely as possible without the initiating of mass movement which would bare the rock surface. If the latter takes place, vegetation reinvades the area, gaining an easy foothold on the weathered rock, and tends to inhibit further washing or mass movement on the nearly cleared surface.

It thus appears that the movement of mantle rock fragments by wash or undermining takes place on gentler slopes than does ordinary

FIGURE 1. CHARACTERISTIC WALL OF LARGER VALLEY
With forested upper slopes of 35 to 70 degrees cut on rock above and gentler slopes cut on weathered talus fill below. (Kalihi Valley).

FIGURE 2. ONE OF THE LONGER SLIDES
500 feet in vertical extent, seen from opposite wall. (Lulumahu Valley).

FIGURE 3. DETAIL OF LOWER PORTION OF SOIL AVALANCHE
Where some lodging of material has taken place.

VALLEY WALL, SLIDE, AND DETAIL OF SLIDE SURFACE

FIGURE 1. DETAIL OF UPPER PART OF GRASSY SLOPE OF ONE OF LARGER VALLEYS
Showing several slide scars. More distant scar exposes weathered bedrock, those in foreground show rudely stratified colluvium. (Moanalua Valley).

FIGURE 2. DETAIL OF STRIATED SURFACE OF WEATHERED BEDROCK
Exposed in slide scar. (Palolo Valley).

SLIDES ON VALLEY WALL AND DETAIL OF SLIDE-STRIATED SURFACE

mass slippage. Vegetation inhibits the movement of debris and interferes with the maintenance of the lower slope. With continued addition of material from above, slopes tend to be oversteepened, so far as colluvial forces are concerned, and the adjustment by slippage is resisted longer by root systems. Thus the process becomes slightly more discontinuous in its action, and the adjustment by sliding slightly more catastrophic than would otherwise be.

FORMATION OF KNIFE-EDGE RIDGES

The most active sliding is in the axial and upper portions of chutes, which head up to the saddle parts and more characteristically knife-edged parts of interfluve ridges. Sliding on the surfaces of ribs is toward the chutes and does not so effectively reduce the slopes of the axes of the ribs to the typical uniformity or angle of slope of slide-determined surfaces. Ridges tend to alternate between the more rounded, higher knobs which lie opposite the ribs and the lower saddles which are more often developed to narrow knife-edged form.

Since the upper parts of chutes are cut more actively by sliding than by ordinary stream activity, the narrowing of the ridges at the chute head is due chiefly to the sliding process. This would be true even if the characteristic slope of the slides continued straight up to the ridge. However, in addition it is most common for the slide to end on the upslope side in a nearly vertical, or even overhanging, cliff, formed where the sliding material has broken away from the better rooted vegetation and more stable rock which forms the ridge crest (Fig. 2). The greater stability of material on the crest is due to its balance on the ridge summit and to the tendency for a more stable root system and larger trees to develop in this less disturbed situation than on the immediately subjacent and steeper slopes.

The junction between material sliding on a slope, downward away from material retained on the same slope higher up, tends to be vertical, since, under the conditions postulated, the upper parts of a slide tend initially to ride on the lower parts, and the line between moving and nonmoving substratum would tend to be projected upward vertically in response to gravity, or even outward normal to the slope (Fig. 2). Hence, insofar as the material can stand in vertical banks, and subject to the modifying effect of roots, the stratified character of partly weathered lava flows, the edges of material left on a ridge (or elsewhere on a slope) and from which slides have moved downslope tend to be nearly vertical.

In most places the crests of ridges are composed of weathered, nearly structureless residual material, or claylike soil, but in others the crest is

composed of residual blocks of weathered lava flows in place. Both types are commonly steep-sided and may slope away for the first 50 feet at angles of 60 to 85 degrees. Below, the slope commonly is reduced to that normal to the slide surfaces. It seems clear that the narrow ridge

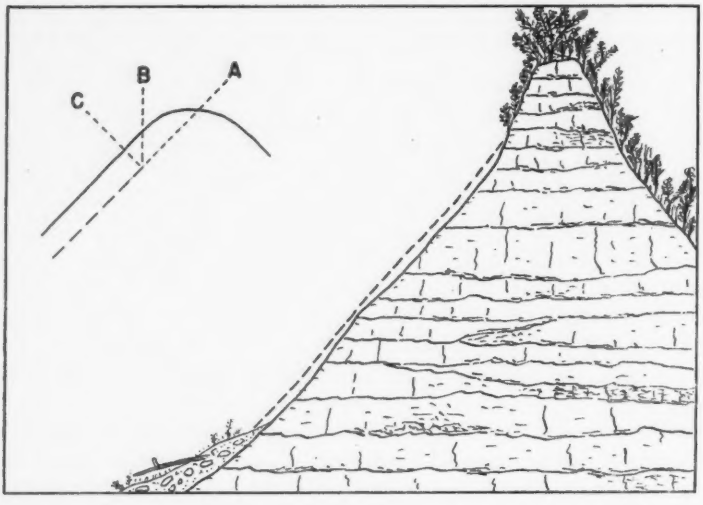

FIGURE 2.—*Diagram showing probable function of soil avalanches in developing knife-edge ridges*

Upper limitation of the moving material involved in the slide is more likely to follow line B than A and for 1 or 2 feet may follow line C. Net result is development of narrow ridge top with immediate slopes even steeper than those lower down.

crests are commonly residual masses from which, by successive sliding, the material has broken away on either side, since many examples of fresh slides, or somewhat older but identifiable slides, broken away from ridges have been seen.

CYCLE OF ACTIVITY

To assume the significance postulated, slides must be recurrent. It is not necessary, however, that they recur in one place with great rapidity. If 1 foot be removed from one slide area and if the next slide be assumed to consist of 1 foot of nearly loosened decomposed rock from beneath it, a slide every 1000 years in each spot would be rapid erosion. In checking over possible rates, it is thought that in the past 8 years, in an area of about 15 square miles, in the part of the Honolulu watershed best known to the writer, there have been the equivalent of 200 slides

involving an acre each. If it can be assumed that the material removed is eventually carried downslope and to the ocean by streams, the removal of 1 foot of average thickness from these areas is equivalent to a rate of about 1 foot in 400 years for the whole area. Even if these assumptions favor too high a rate, the process can be very significant in land reduction.

We may properly commence with an area recently bared by sliding. Some areas reveal 25 to 50 per cent of rock outcrop, in which plants might find difficulty in rooting; others are at least veneered by soil left after the sliding. However, practically no slide scars of this type have been seen on which plants do not advance rather readily. Scars 6 months old are still somewhat brown as seen from a distance; those 3 or 4 years old are green and are identified by lack of trees and the contrast seen at their boundaries.

Very little is known of the plant succession on such scars, and doubtless the rooting and propagation habits of these plants are important factors in the sliding. Development of a somewhat loosened layer, favorable for sliding, is no doubt due to a variety of factors. Probably the root systems of plants serve to bind the upper soil mass together and yet to render it somewhat more porous and subject to movement. Some finer-grained detrital material is added by colluvial wash from above. Absorbed water increases the weight of the loosened zone. When the process of loosening, of integration of the mass through growth of plant roots has gone far enough, eventually under stress of excess water saturation the whole mass will commence to slide.

On most slides observed, there were few large trees. The relation may be reciprocal; growth of large trees may inhibit slides, but large trees could not develop on a surface where sliding is active. Slides are probably a factor in the invasion of staghorn ferns and other introduced plants which are detrimental to the native forest. The wild pig is considered a menace by his disturbing of the plant cover. Without dissenting from this view, it appears to the present writer that these shallow soil avalanches are at least of equal importance in the opportunity they afford for the invasion of undesirable plant pioneers, and they represent a factor which must have been operative since time immemorial. Indeed, they are entirely a normal process which has probably been little affected by human activities. Whether prompt attention to such slides by planting desirable seedlings can in some measure combat the invasion of detrimental vegetation is a question the writer is not able to answer.

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CRETACEOUS-TERTIARY BOUNDARY IN THE DENVER BASIN, COLORADO

BY ROLAND W. BROWN

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ABSTRACT

Field work by the writer during 1939, 1940, and 1941 resulted in the discovery of Paleocene mammals only 50 feet above the highest remains of *Triceratops* in the Denver Basin, Colorado. This new evidence checks that of the floras, and thus indicates within relatively narrow limits the position of the Cretaceous-Tertiary boundary in that area. As this boundary definitely places the basal portion of the Denver formation and Dawson arkose in the Upper Cretaceous sequence, some changes in the nomenclature of these and other strata adjacent to the boundary become necessary. Stratigraphic and paleontologic evidence shows that the Laramie formation, Arapahoe formation, and the Cretaceous parts of the Denver formation and Dawson arkose comprise a unit correlative with the Lance formation and its equivalents. Therefore, the writer proposes that the Laramie be redefined to include all the Upper Cretaceous sequence between the top of the Fox Hills and the base of the Paleocene in the Denver Basin; that the term Arapahoe as Arapahoe conglomerate member be retained for the conglomerate immediately overlying the present Laramie formation; and that the terms Denver formation and Dawson arkose be restricted to the Tertiary strata.

INTRODUCTION

For the past 50 years the position of the Cretaceous-Tertiary boundary in the Denver Basin, Colorado, has been a moot question. This problem, however, is not peculiar to the Denver Basin but involves a large portion of the Rocky Mountain and Plains region. Differing chiefly in their interpretations of the fossil record, various students have assumed the boundary to be at widely separated stratigraphic horizons. Knowlton (1922, p. 3-81) reviewed the situation at length and concluded that the boundary in the Denver Basin is at the contact of the Laramie and Arapahoe formations. Few geologists and paleontologists accepted this conclusion as final.

Since 1929 the writer has been searching the Rocky Mountain and Plains region for geologic and paleontologic information that might solve the boundary problem for the entire area. He accumulated much new plant evidence and during the field seasons of 1939, 1940, and 1941 found mammalian remains at critical localities in the Denver Basin hitherto yielding only fossil plants. The description of these significant fossil mammals has already been published by Gazin (1941). The present paper reviews the pertinent geology in some detail and suggests the practical implications of the recent discoveries for possible changes in the nomenclature of the strata adjacent to the boundary line.

FORMATIONS INVOLVED

The strata of the Denver Basin particularly concerned in discussions of the Cretaceous-Tertiary boundary question are shown in Table 1. These strata are best exposed on the rim of a roughly oval, synclinal basin (Fig. 1) extending from the vicinity of Brighton to Colorado

TABLE 1.—*Some Upper Cretaceous and Tertiary formations in the Denver Basin, Colorado*

North	South
Denver formation Arenaceous clays, shales, conglomerates with andesitic debris, basalt flows. 400–1200 feet.	Dawson arkose Andesitic conglomerates, clays, shales, coals, arkosic conglomerates and sandstones, tuff, rhyolite flows. 2000 ± feet.
Arapahoe formation Conglomerate, arenaceous clays. 500 ± feet.	
Laramie formation Sandstones, shales, clays, coals. 600–1200 feet.	Laramie formation Sandstones, shales, clays, coals. 300 ± feet.

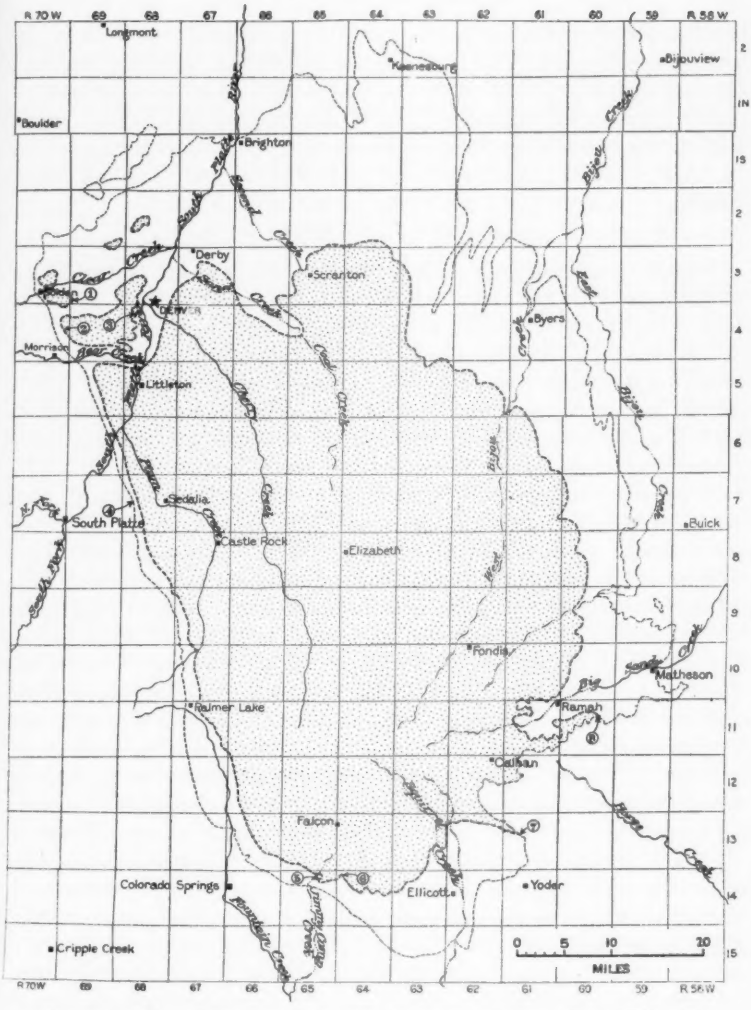


FIGURE 1.—*Map of Cretaceous-Tertiary boundary in Denver Basin, Colorado*

Light dashed line is lowest exposure of Arapahoe, Denver and Dawson strata, as adapted from the 1935 Geologic Map of Colorado. Heavy dashed line surrounding stippled area is Cretaceous-Tertiary boundary as interpreted by the writer.

(1) Mammal locality on South Table Mountain; (2) Section on west slope of Green Mountain; (3) Clay pit of Overland Pressed Brick Company; (4) Douglas (Lehigh) coal mine; (5) Mammal locality at west end of Corral Bluffs; (6) Mammal localities near east end of Corral Bluffs; (7), (8) Cretaceous-Tertiary boundary disappears beneath cover of late Tertiary and Pleistocene deposits.

Springs and from the Rocky Mountain front to points as much as 50 miles eastward on the Plains. Along the foothills the beds are upturned, but they flatten rather abruptly toward the east or northeast. A number of geologists have described portions of this Basin as will be cited at the appropriate places. The latest additions, reviews, or comments have been by Lovering (1929), Dane and Pierce (1936), and Van Tuyl and others (1938).

HISTORY OF THE PROBLEM

In recent years students of the strata and fossils of the Denver Basin have sought the Cretaceous-Tertiary boundary somewhere in the basal part of the Denver formation as defined by Cross.[1] Originally, however, the striking lithologic discontinuity at the contact of the Laramie and overlying Arapahoe formation (Fig. 1) received most attention. This contact is exposed in vertical position near Golden in a clay pit south of the athletic field of the Colorado School of Mines; in an abandoned clay pit near 20th and Illinois Street; and half a mile southwest of the Industrial School. No angular discordance is evident at these exposures, but lenses of clay and sand alternate with thin conglomerates before the main Arapahoe conglomerate appears. On the southeast side of the Denver Basin the contact is apparently at the horizon where drab shale of the Laramie is overlain by lenses of cross-bedded arkosic conglomerate and sandstone (Dane and Pierce, 1936, p. 1320, 1321). Although the contact of the Arapahoe conglomerate with the clays and shales of the Laramie appears to be erosional and transitional rather than angular at these localities, Emmons, Cross, and Eldridge (1896) believed it represented an unconformity denoting a great lapse of time. Some commentators, however, seem to have misunderstood and thereby to have exaggerated the original opinion of these authors by mistaking "exposed to erosion the upturned edges of 14,000 feet of strata" to mean "removal of 14,000 feet of strata over the entire area" (Stanton, 1914, p. 347; Lovering, 1929, p. 92; Dane and Pierce, 1936, p. 1325). The latter interpretation, despite its implication by Cross (Emmons, *et al.*, 1896, p. 207), does not seem to have been the unanimous actual intent of Cross and his colleagues, for he (p. 200) states the immediate pre-Arapahoe situation clearly, and Emmons (1896, p. 29, 31) says plainly:

"In this post-Laramie movement not only was there a general continental elevation of the whole region, but the mountain areas suffered a differential uplift in relation

[1] In 1935 the United States Geological Survey, recognizing the evidence of the dinosaur and plant fossils, accepted the age of the Denver formation as Upper Cretaceous and Eocene (?). With the revision in 1939 of the Survey's classification to recognize Paleocene as a series distinct from Eocene, and upon the evidence presented in this and Gazin's (1941) paper, the 1935 designation has now been changed to Upper Cretaceous and Paleocene. Further changes in nomenclature proposed here have not yet been adopted by the Survey.

to the surrounding plains or lowlands, so that the edges of the strata resting against these flanks were in many places upturned at considerable angles. . . . That the Mesozoic strata were uplifted along the foothills above the general level of the beds, and their upturned edges exposed to erosion, is proved by the fact that rolled fragments of the rocks of the different formations are found in the succeeding Arapahoe and Denver formations. . . . After an erosion of the Laramie beds which removed from portions of the Denver Basin 600 feet or more of the previously deposited sediments, . . . sedimentation again set in."

A pre-Arapahoe erosion of the Laramie to a depth of 600 feet, locally, may be a close approximation, but the few good exposures of the Laramie-Arapahoe contact seem inconclusive on this point.

Emmons, Cross, and Eldridge (1896, p. 248-252) did not definitely date the Arapahoe and Denver formations because they had misgivings aroused by the alleged inconsistency of having a dinosaur fauna, considered by the paleozoologists to indicate Upper Cretaceous age, in strata containing a flora said by the paleobotanists to indicate Tertiary age. Later, however, Cross (1907) and others strongly urged the reference of the Arapahoe and Denver to the Tertiary.

Much of the ensuing discussion of this problem had a basis of insufficient information. For example, in the Laramie formation fresh and brackish-water mollusks and a sizable flora indicated an Upper Cretaceous age. Lack of authentic evidence of dinosaurs, however, caused Knowlton (1911) to inquire, Where are the Laramie dinosaurs? The point he sought to make was that in his opinion the Laramie beds of the Denver Basin, showing little or no evidence of dinosaurs, are not the equivalents of the noted dinosaur-bearing Lance beds in eastern Wyoming or of other dinosaur-bearing strata said to lie unconformably upon the Medicine Bow formation near the mouth of Medicine Bow River in Wyoming. He insisted, therefore, that the beds immediately overlying the Laramie are considerably younger and are most probably equivalents of the Lance, which, after the lapse of a decade, he now believed were Tertiary—a change of belief that caused much of the later Laramie-Lance-Fort Union confusion. Nevertheless, he concluded his paper open-mindedly: "If there is valid evidence to show that the Lance formation ('*Ceratops* beds') is the equivalent of the Laramie in whole or in any part it would be welcome." Knowlton to the last contended that the Lance should be included in the Tertiary, but his question, Where are the Laramie dinosaurs? was answered by the finding of horn cores of *Triceratops* in Laramie beds near Briggsdale, Colorado (Toepelman, 1926). In 1940 the writer obtained from Mr. George W. Prince, of Morrison, Colorado, two horn cores of *Triceratops* which Prince had collected from his clay pit in the Laramie on Mount Carbon, on the south side of Bear Creek, 3 miles east of Morrison.

Almost from the beginning of geologic work in the Denver Basin dinosaurs were found in both the Arapahoe and lower part of the Denver formation. The writer visited many of the old localities and found additional fragments. He also found new localities in the Arapahoe conglomerate near Golden; in the basal strata of the Denver formation exposed in the lower levels of the clay pit of the Overland Pressed Brick Company on the west side of the South Platte River at Denver; in the lower part of the Dawson arkose on Jimmy Camp Creek, 9 miles east of Colorado Springs; and on Black Squirrel Creek, northeast of Ellicott, where Dane and Pierce previously found fragments.

Remains of *Triceratops*, therefore, occur indigenously from the Laramie up into the lower strata of the Denver formation. Thus, most of those geologists who accepted the dinosaurian evidence regarded the Denver and Arapahoe formations as Upper Cretaceous and placed the Cretaceous-Tertiary boundary at the top of the Denver formation. Others, because of the supposed great unconformity between the Laramie and Arapahoe and because of a large Tertiary flora from horizons just above the occurrences of the Denver dinosaurs, insisted that both the Denver and the underlying Arapahoe must be Tertiary, and that the boundary is at the Laramie-Arapahoe contact.

By contrast with this ambiguous situation in the Denver Basin that on Lance Creek in eastern Wyoming was for many years relatively clear. The Cretaceous and Tertiary there had been correctly separated in 1897 by recognition of the differences between their respective faunas and floras (Stanton and Knowlton, 1897). Twenty-five years later, Knowlton (1922, p. 28), reviewing that work, stated how the fossil plant evidence aided in effecting the separation:

> "The plants [of the *Ceratops* beds], of which 25 forms were listed, were compared with the floras of various localities then supposed to be of Laramie age, and the conclusion was reached that 'as nearly as can be made out, the plants confirm the Laramie age of the *Ceratops* beds'."
>
> "From the beds conformably overlying the highest dinosaur-bearing stratum 18 species of plants were obtained, and of the 10 named species 9 proved to be typical Fort Union forms, and hence these beds were regarded as of Fort Union age."

Soon after the classification of the coal lands began, attempts—on the basis of similarity in color and lithology—to correlate other strata in Wyoming, Montana, and the Dakotas with those on Lance Creek led to inclusion in the Lance of beds now dated as Tertiary on paleontologic and stratigraphic evidence. A spurious "Upper Lance" thus came to be recognized and erroneously used even to the present time. It should be emphasized that the term Lance as originally used for the type locality applied only to the dinosaur-bearing strata.

A restudy of the original plant collections involved in this boundary question, plus new collections, indicates fairly definitely that no Tertiary flora existed contemporaneously with triceratopsian dinosaurs. At numerous localities in the Rocky Mountain region the writer has observed that remains of the latest dinosaurs underlie sandy, relatively barren deposits, sometimes 50 feet or more thick, the latter overlain by lithologically different beds containing a Paleocene fauna and flora. Seemingly these barren strata, otherwise not significantly different from the underlying beds, should yield dinosaur remains if dinosaurs had existed at the time of their deposition. During this interval the fauna and flora of the Rocky Mountain and Plains region changed materially. The immediate progenitors of the Paleocene flora may have left remains in the Upper Cretaceous, but study of these floras reveals that surprisingly few Cretaceous species maintained their identity into the Paleocene (Dorf, 1940, p. 217; 1942, p. 120). Probably this number will be increased when better collections are found near the contact. Apparently, at the close of Upper Cretaceous time physical and climatic changes in the Rocky Mountain region created conditions that prohibited the survival in that area of much Upper Cretaceous life. The resulting emigration, immigration, or rapid evolution of species so changed the fauna and flora that these groups are now frequently referred to as having a "Tertiary aspect"—that is, one beginning to exhibit a recognizable modern accent.

The dinosaur-bearing strata of the Denver formation and Dawson arkose are generally sandy and include few shaly beds that yield identifiable fossil plants. The writer and others, however, have made several small collections, as follows: in the lower levels of the clay pit of the Overland Pressed Brick Company at Denver; in the Dawson arkose on the east bank of Jimmy Camp Creek, about 600 feet north of the bridge on State Highway 94; and at several other localities in equivalent strata but without dinosaurs.

The citation of these localities necessitates a review of some opinions concerning the stratigraphic levels from which a few of the older collections were obtained. In the description of the flora of the Denver and associated formations Knowlton (1930, p. 6) considered two collections supposed to have come from the Arapahoe formation or its equivalents: on Sand Creek, about 15 miles east of Denver; and the vicinity of the old Douglas coal mine, 4 miles west of Sedalia. He removed the first from the Arapahoe and referred it to the Denver on the basis of a well log reported by W. T. Lee (Knowlton, 1922, p. 103). The location of the well is given as sec. 24, T. 3 S., R. 67 W. As will be shown later, the

Arapahoe and Denver strata containing dinosaur remains near Golden, are approximately 800 feet thick. Knowlton's interpretation of the well log indicates a thickness of 850 feet for this unit. If little or no thinning of the Arapahoe and dinosaur-bearing Denver unit occurs between Golden and the site of the well, and if Knowlton's interpretation of the well log be accepted, the top of that unit, or, in other words, the Cretaceous-Tertiary boundary, should not be far beneath the surface at the site of the well. However, this well is on the gently westward-dipping flank of a low anticline, the crest of which lies between the well and Scranton, because at the abandoned Scranton coal shaft the strata are said to dip 3° to 5° eastward (Emmons, *et al.*, 1896, p. 373). The plant locality on Sand Creek is 8 miles southeast of the well and topographically 175 feet higher, but it is apparently on the crest of the fold, thus canceling any greater thickness the section there might have over that at the well. The probability, therefore, that the Sand Creek locality is high in the dinosaur-bearing Arapahoe-Denver unit is heightened by the fact that one of the few identifiable plants found there is *Eriocaulon? porosum* Lesquereux, which is synonymous with *Paleoaster inquirenda* Knowlton, a characteristic but botanically undetermined species, which, so far as the writer is aware, has never been found in strata younger than Upper Cretaceous. A similar specimen, called, however, *Sterculiocarpus coloradensis* Berry, was found at approximately the same stratigraphic level high in the Upper Cretaceous portion of the Dawson arkose but beneath the workable coal seam cropping out in the valley of Big Sandy Creek, near Ramah, in the southeast quarter of the Denver Basin.

Fossil plants have been collected at and above the coal seam at Ramah and have been identified as Tertiary because many of the species are identical with those from the Tertiary Denver in the northern part of the Basin, and also because a number of these species have never been found in the Upper Cretaceous. The writer has traced this coal zone northwestward to a point 10 miles north of Fondis, 30 miles short of the Scranton coal field. Richardson (1917, p. 243, 244) assumed that the coal in both areas is in the same zone and that the Scranton coal is Tertiary. No distinctive fossil plants have been collected from the decayed dumps of the Scranton mine, and Emmons, *et al.* (1896), although recording a section of the Scranton coals, do not give the depth of the coal section beneath the surface. Water and oil well logs from scattered points in the Scranton field do not seem altogether reliable. Consequently, the correlation of the Scranton coals with those near Ramah must await satisfactory fossil collections, better subsurface records, or accurate lateral tracing.

Knowlton retained the second or Sedalia locality in the Arapahoe. In 1938 the writer, with W. T. Lambert, owner of the ranch on which the now abandoned mine is located, visited the area and measured the section.

SECTION AT DOUGLAS (LEHIGH) MINE

Sec. 20, T. 7 S., R. 68 W., west of Sedalia, Colorado.

	Feet
9. Arkosic beds	25+
8. Arthur Lakes' plant localities, 1900 and 3000 feet E. of coal mine	50
7. Arkosic beds, nearly horizontal	50
6. Covered	150
5. Reddish and dark shale, and light-colored arkosic conglomerate, dip 45° E.	125
4. Covered	700
3. Arapahoe conglomerate, dip 80° E.	30
2. Sandstone and shale	500
1. Coal, dip 82° E.	11
Total	1641

At least 1000 feet of strata lie between the top of the Laramie (no. 2) and the level of the plant localities (no. 8). This thickness, greater than any accepted for the Arapahoe formation, clearly indicates that the plant localities are not in the Arapahoe formation but in that part of the Dawson arkose which is laterally equivalent to the Tertiary Denver. The fossil plants amply confirm this conclusion.

SPECIFIC LOCALITIES

Inasmuch as the phenomena at the Laramie-Arapahoe contact seemed to represent only an episode in Upper Cretaceous history, and because hundreds of feet of sediment carrying remains of *Triceratops* and a distinctive flora were deposited upon the Laramie, the writer concluded that the Cretaceous-Tertiary boundary should be sought higher in the section but beneath the horizons yielding the Tertiary Denver flora. Accordingly, he concentrated attention on that basal part of the Denver formation which more particularly overlapped in the conflicting claims of the paleozoologists and paleobotanists. Available outcrops of this zone are not numerous, and many are barren. On June 13, 1939, however, on the bare slope (Pl. 1, figs. 1, 2) on the southeastern flank of South Table Mountain, about 3 miles east of Golden, in the SW ¼ NW ¼ sec. 31, T. 3 S., R. 69 W,. the writer found the highest dinosaur remains in drab-green, sandy beds near the foot of a ravine and 50 feet higher in light-colored sandy clay, some small mammal jaws and teeth, together with fragments of turtles and crocodiles. Additional mammal specimens were found in 1940. The best specimen, since described as *Baioconodon denverensis* Gazin (Pl. 2, figs. 9, 11), indicates Paleocene age. The typical Tertiary Denver flora occurs with the mammals and at intervals from this level to the base of the lava cap 170 feet above.

The geologic section at this locality includes the following conformable, nearly horizontal strata:

SECTION ON SOUTHEASTERN SLOPE OF SOUTH TABLE MOUNTAIN, GOLDEN, COLORADO

	Feet
10. Dark basaltic lava at top of mesa	10
9. Drab-green, andesitic sandstone, carbonaceous shale, thin conglomerates, fossil plants	70
8. Gray sandstones with thin dark shales	75
7. Light-colored, arenaceous clay, mammals, turtles, crocodiles, leaves, fossil wood	25
6. Coarse brown sandstone	2
5. Sandy shale tinged with pink and green	11
4. Drab-green, sandy shale tinged with lavender	20
3. Rusty yellow sandstone	14
2. Cross-bedded, drab conglomerate	3
1. Drab-green andesitic sandstone, with dinosaur and turtle bones	21
Foot of ravine at fence — Total	251

From a distance (Pl. 1, fig. 1) this section appears to have three divisions: a lower dark zone, a middle light-colored zone, and an upper dark zone. The mammals were found near the base of the middle zone. Apparently, therefore, the Cretaceous-Tertiary contact lies within the limits of the 50-foot interval between the dinosaur-bearing sandstone (no. 1) and the light-colored clay with mammals (no. 7). As no fossils were found in this interval, the writer, considering the conspicuous change in lithology at the base of the light-colored beds (no. 7), regards that horizon as the most acceptable position for the boundary.

Cross (Emmons, *et al.*, 1896, p. 185) originally included this entire section in the Denver formation and estimated the stratigraphic interval between the basalt and the top of the Arapahoe to be from 450 to 475 feet. Therefore, as the base of the mammal-bearing bed is 170 feet below the basalt, about 290 feet of the lower part of the Denver formation in the vicinity of Golden must be included in the Upper Cretaceous sequence. If to this be added 500 feet, the estimated average thickness of the underlying Arapahoe formation, the total thickness of the Arapahoe-Cretaceous Denver approximates 800 feet.

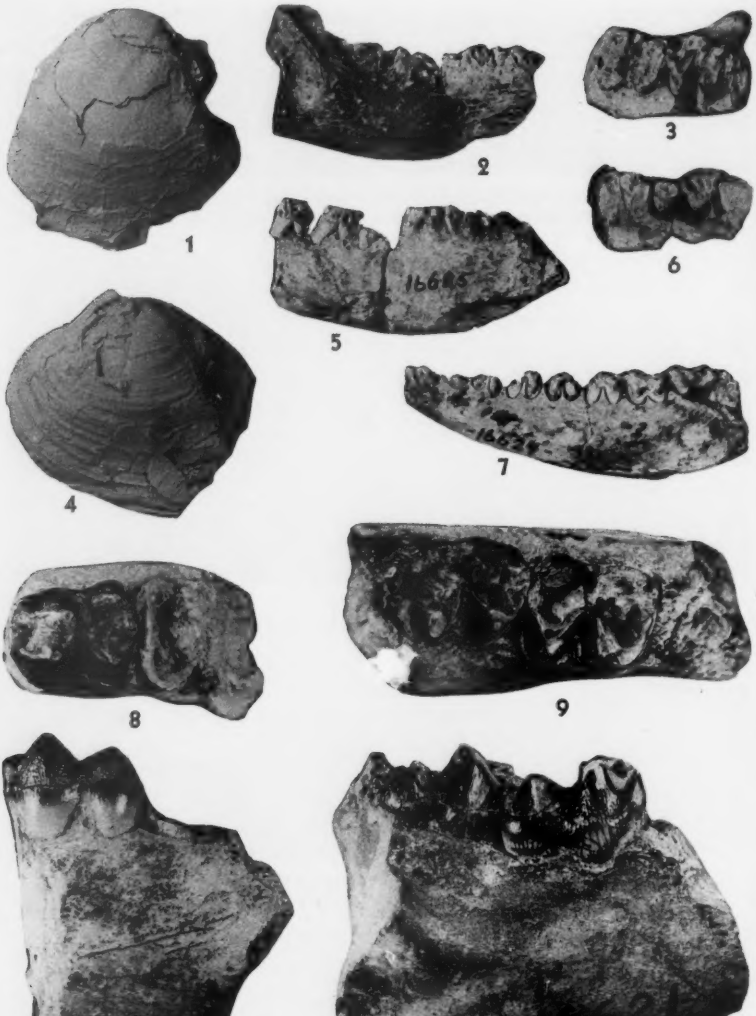
Talus on much of the slopes of the Table Mountains conceals the Cretaceous-Tertiary contact. The west slope of Green Mountain (Pl. 1, fig. 3), about 4½ miles southeast of Golden, displays a section of Arapahoe and Denver which has been studied by a number of observers, including Cross (Emmons, *et al.*, 1896, p. 171-179), John D. Marr (Waldschmidt, 1939, p. 42-44), and the writer. As no diagnostic fossils have been found at this locality the position of the boundary must be inferred. The writer places it at the base of Marr's 82-foot light-colored unit of "shale, gray and sandy with fossil wood near the top" in his "Basalt-Andesite Zone." Marr's unit, perhaps significantly, yields fossil wood, as does also the mammal-bearing unit on South Table Mountain.

FIGURE 1. SOUTHEASTERN SLOPE OF SOUTH TABLE MOUNTAIN
Mammal locality as seen from U. S. Highway 40, east of Golden.

FIGURE 2. DETAIL OF FIGURE 1
Man stands where mammal jaw was found in Denver formation. Arrows indicate Cretaceous-Tertiary boundary.

FIGURE 3. WEST SLOPE OF GREEN MOUNTAIN
Arrow indicates probable position of Cretaceous-Tertiary boundary.

FIGURE 4. SPUR OF CORRAL BLUFFS
Arrow indicates where mammal jaw was found in Dawson arkose.



(1) *Corbicula cleburni* White; (4) *Corbicula planumbona* (Meek); (3) (5) (6) *Conacodon* cf. *entoconus* Cope; (2) (7) (9) (11) *Baioconodon denverensis* Gazin; (9) top view, and (11) side view of the type, x 3; (8) (10) a creodont, near *Eoconodon heilprinianus* (Cope), x 2.

UPPER CRETACEOUS INVERTEBRATES AND PALEOCENE MAMMALS FROM DENVER BASIN, COLORADO

A second locality at which Paleocene mammals were found in the Denver Basin is in the area east of Jimmy Camp Creek, which is crossed 9 miles east of Colorado Springs by State Highway 94. From a point 2½ miles east of the bridge crossing, the region north of the highway appears as a huge amphitheater backed by a wall of badlands called Corral Bluffs. To the east the highway rises to a deep road cut through the crest of the bluffs. These bluffs are capped by the coarse upper part of the Dawson arkose. Beneath are the somber shales, sandstones, and coals of the lower part of the Dawson. The mammal *Conacodon* cf. *entoconus* (Cope) (Pl. 2, figs. 3, 5, 6) was found in 1940 in the SW ¼ NE ¼ sec. 6, T. 14 S., R. 64 W., in the northeast quarter of the amphitheater, on the saddle of a long spur (Pl. 1, fig. 4) projecting southward toward the highway, approximately 300 feet above the level of the highway and about 200 feet below the general level of the plain at the top of the bluffs. The horizon is evidently high in the lower part of the Dawson. A portion of the jaw of *Carsioptychus* sp. was found in 1941 half a mile farther west at about the same level; and in 1932, at a level probably 200 feet lower and 1 mile south, C. Lewis Gazin and L. W. Nicklaus found part of a jaw, with one perfect molar, of a Paleocene creodont near *Eoconodon heilprinianus* (Cope) (Pl. 2, figs. 8, 10).

One other mammalian fragment from this area needs to be mentioned because it has literally been a "bone of contention." Richardson (1912, p. 272; 1915, p. 8) reporting the finding of this bone, which Gidley identified as the tibia of an Eocene creodont, gives the locality as the SW ¼ sec. 2, T. 14 S., R. 65 W., 600 feet above the base of the Dawson arkose. This locality, therefore, is about 2 miles west of the 1940 and 1941 locality just described and overlooks the valley of Jimmy Camp Creek to the north. In 1912 W. T. Lee visited this area and found dinosaur bones in sec. 3, T. 14 S., R. 65 W., on the south side of Jimmy Camp Creek, about 500 feet above the base of the Dawson arkose. He thereupon published a paper describing the purported discovery of dinosaurs in the Tertiary (1913, p. 531).

Here was a renewal of the suggestion that some Cretaceous and Tertiary species, which orthodox paleontologists have always considered as separated in time, may be contemporaneous. To test the validity of Lee's conclusion it would seem necessary to (1) check the bones as to correct identification; (2) visit the area to check localities and stratigraphic levels. It is now apparently impossible to check the bones because they seem to be lost. As to checking localities, the writer has examined the area, measured the geologic section, and collected fossils with the following results: (1) The section, beginning at the highway bridge over Jimmy Camp Creek and proceeding northeastward 1 mile

to the top of the ridge, includes more than 600 feet of northeastward-dipping strata from dark, andesitic conglomerates, sandstones, and shales (Goldman, 1910, p. 300; Finlay, 1916, p. 9), dipping 10°, to coarse, white, arkosic beds, dipping less than 5°. (2) A Cretaceous flora and dinosaur bones occur in the lower 500 feet of this section, the bones being rather abundant in the dark or somber sandy and shaly beds exposed on the south side of Jimmy Camp Creek a quarter to half a mile east of the Richfield Springs (now Banning-Lewis) ranch buildings. (3) No dinosaur bones occur above this dark band of strata. (4) The creodont tibia was found on a hill in section 2 above horizons carrying typical Tertiary Denver plants, but no dinosaur-bearing beds crop out in that section. Lee's dinosaur bones came from section 3, 100 feet topographically lower than the creodont tibia. Lee, therefore, confused two sites and two levels. This conclusion is confirmed by a quotation from Knowlton's field notebook for July 21, 1910, the date on which he and Richardson examined this area: "A further indication of possible Denver [that is, Tertiary] age is the finding by Richardson of turtle plates and a small bone in andesitic beds 100 feet higher than the highest plants."

The plants referred to by Knowlton are from his locality 5 and are unquestionably the same species as those from the mammal beds at Golden. Knowlton's locality 5 is half a mile farther east and 100 feet topographically higher than his locality 1, which is half a mile east of the Banning-Lewis ranch buildings. At locality 1 he collected Upper Cretaceous plants and near there Lee most probably got his dinosaur bones. Evidently, the Cretaceous-Tertiary boundary here, as at Golden, separates Cretaceous dinosaurs from diagnostic Tertiary animals and plants.

No dinosaurian remains have ever been reported from above the zone of workable coals in the vicinity of Ramah in the southeastern part of the Denver Basin, east of the Corral Bluffs area. At and above this horizon, however, occurs the typical Paleocene flora similar to that in and above the mammal beds on South Table Mountain at Golden.

The evidence, therefore, from the fossil plants and associated mammalian remains permits limitation of the general position of the Cretaceous-Tertiary boundary in the Denver Basin, although the actual boundary is not sharp and cannot be traced with absolute certainty around the entire Basin. Between Ramah and Scranton, on the east side of the Basin, outcrops are few and poor and the boundary is concealed. Its position as shown on the map (Fig. 1) should therefore be regarded as tentative. Few formational boundaries as mapped in these

continental beds are clearly defined at all outcrops (Cross, *in* Emmons, *et al.*, 1896, p. 159; Dane and Pierce, 1936, p. 1313, 1320).

The mammal-bearing portions of the Denver formation and Dawson arkose are Paleocene and belong to the Puercan or oldest division of the Tertiary. They are the equivalents of the lower part of the Fort Union sequence in Wyoming, Montana, and North Dakota and South Dakota and of strata now known by other names in Colorado, New Mexico, and Utah. The overlying coarser and lighter-colored strata at the surface of the plateau region between the South Platte and Arkansas River drainage were originally called the Monument Creek group. The upper part of this group was subsequently differentiated as the Castle Rock conglomerate of Oligocene age, whereas the lower part was included in the Dawson arkose. This phase of the Dawson arkose has been depicted as unconformably overlying Laramie, Arapahoe, and Denver strata (Emmons, *et al.*, 1896, p. 195-199). Few fossils have been found in these beds, but G. B. Richardson and the writer independently obtained small collections of plants from a sandy lens in the lower strata in the railroad cut just west of Falcon, northeast of Colorado Springs. Knowlton identified Richardson's collection and suggested that it might indicate an age as young as middle Eocene. Richardson's and the writer's collection comprise an assemblage that can be matched readily with species from the Paleocene and earliest Eocene of Colorado and adjacent eastern Wyoming. The so-called upper Dawson may consequently not be very much younger than the underlying mammal-bearing beds, despite the local evidence of an erosion interval.

SUGGESTED CHANGES IN NOMENCLATURE

This determination of the position of the Cretaceous-Tertiary boundary in the Denver Basin requires some readjustment in the classification of the formations adjacent to the boundary. The changes suggested by the writer are indicated in Table 2. He proposes that the Laramie formation include all the dinosaur-bearing Arapahoe, Denver, and Dawson arkose; that the terms Denver formation and Dawson arkose be applied only to the Tertiary strata; and that the term Arapahoe be restricted to an Arapahoe conglomerate member of the Laramie. Thus the Laramie would comprise all the Upper Cretaceous strata between the top of the Fox Hills and the base of the Paleocene in the Denver Basin.

Although it is not a new proposal (Johnson, 1931, p. 370), the elimination of the term Arapahoe as a formation name may be questioned as desirable. However, the Arapahoe formation is not a satisfactory mappable unit, because its upper limits—the first appearance of andesitic

debris being said to mark the base of the Denver formation—are indefinite, as admitted by Cross (Emmons, *et al.*, 1896, p. 159), who said:

> "The actual contact of Denver and Arapahoe beds is seldom well exposed, hence a complete description of the change from one to the other can not be given. The boundary line of the Denver formation as drawn upon the map is approximate, being clearly established at but few points."

Furthermore, in the valley of Cherry Creek, a few miles southeast of Denver, the Arapahoe-Denver deposits interfinger with the Dawson

TABLE 2.—*Present and proposed classification of formations adjacent to Cretaceous-Tertiary boundary in Denver Basin, Colorado*

	Present		Proposed		
	North	South	North	South	
Upper Cretaceous and Paleocene	Denver formation	Dawson	Denver formation	Dawson arkose	Paleocene
Upper Cretaceous	Arapahoe formation	arkose	Laramie Arapahoe conglomerate member	Laramie formation	Upper Cretaceous
	Laramie formation	Laramie formation			
	Fox Hills sandstone	Fox Hills sandstone	Fox Hills sandstone	Fox Hills sandstone	

arkose—a condition noted by Cross (p. 183) and later elaborated by Richardson (1912; 1915). Both Goldman (1910, p. 320) and Findlay (1916, p. 13) reported andesitic debris in the basal beds of the Dawson arkose in the southern part of the Denver Basin east of Colorado Springs. The presence or absence of andesitic material, therefore, has little or no value for discriminating the Arapahoe from the Denver formation, or their equivalents from one another, and thus the only criterion ever alleged for separating these formations is invalid.

Originally the term Laramie was rather loosely applied to all the conformable lignitic strata above the Fox Hills sandstone. In 1910, on the assumption that conformable relations end where conglomerates of the Arapahoe formation overlie the Laramie, the United States Geological Survey restricted its use to the strata between the Fox Hills sandstone and the Arapahoe formation in the Denver Basin. The assumed Laramie-Arapahoe interval, however, has since been generally considered as a

minor erosional irregularity or a transitional contact reflecting merely increased orogeny to the west. As the Lance of eastern Wyoming and Hell Creek of eastern Montana are now considered indivisible stratigraphic units lying between the Fox Hills and Paleocene they are to be correlated with the present Laramie plus whatever lies above the Laramie and beneath the Paleocene. The Laramie, thus enlarged, comprises a unit 1500 to 2100 feet thick.

Stratigraphic and paleontologic grounds justify this enlargement of the Laramie. Marsh (Emmons, 1896, p. 34) proposed that the Laramie should include the Arapahoe and at least part of the Denver, as also did Russell (1930, p. 158), somewhat indirectly, the basis being the presence of the same types of *Triceratops* and other dinosaurs. The flora of the Arapahoe and dinosaur-bearing Denver and Dawson arkose is meager, but the species identified are Laramie or equivalent forms. No mammals have been reported from these strata.

The invertebrate fauna of the Laramie is known best from the prolific localities northeast of Greeley, Colorado. Few invertebrates have been taken from the Arapahoe and Denver formations, but recently Van Tuyl obtained several specimens (Pl. 2, figs. 1, 4) from the excavation for a water well in the lower part of the Denver formation in sec. 1, T. 4 S., R. 70 W., half a mile south of the Camp George West rifle range, near Golden. Reeside identified these as *Corbicula cleburni* White and *C. planumbona* (Meek), both brackish-water forms, indicating that the sea was probably not far to the east during early Denver time. The existence of coal-forming coastal swamps at about the same time in the Scranton coal field, 15 miles east of Denver, corroborate this. *Corbicula cleburni* is known only from the Laramie of northeastern Colorado; and *C. planumbona* only from the Laramie of northeastern Colorado and the Lance of Lance Creek and Goshen Hole, Wyoming. These two specimens alone may not be greatly significant, but together with the dinosaurs and the flora they indicate that the Laramie includes the Arapahoe and basal part of the Denver.

CORRELATIONS

LARAMIE AND LANCE

The Lance formation of eastern Wyoming is apparently a homogeneous and not readily divisible unit lying between the Fox Hills sandstone and the Fort Union formation. Minor erosional irregularities are present, but there is no evidence of a large or significant unconformity within the sequence. The Lance has yielded a relatively large dinosaurian, mammalian, and invertebrate fauna, and a flora of 70 species (Dorf, 1942, p. 97). Consequently, geologists and paleontologists, attempting

to understand the stratigraphy of adjacent regions, have regarded the Lance as a standard of comparison. Thus Stanton and Knowlton (1897, p. 156), comparing the dinosaur-bearing beds at Black Buttes, in the Bitter Creek Valley of Wyoming, with the Lance, said:

"From the facts now available it seems most probable that in Converse [Niobrara] County and in the Bitter Creek valley the time representatives of the Denver and Arapahoe are undifferentiated portions of a continuous series and cannot be separated from the Laramie. The Fort Union beds are apparently distinguishable by means of their flora, and these mark the upper limit of the Laramie in the areas in question."

Again, Stanton (1909, p. 293), discussing the Upper Cretaceous dinosaur-bearing beds of Wyoming and Montana, said:

"The 'Ceratops beds' are of Cretaceous age as decided by stratigraphic relations, by the pronounced Mesozoic character of the vertebrate fauna with absence of all Tertiary types, and by the close relations of its invertebrate fauna with the Cretaceous. The relations of the flora with Eocene floras is believed to be less important than this faunal and stratigraphic evidence. Taken in their whole extent they probably include equivalents of the Laramie, Arapahoe, and Denver formations of the Denver Basin."

In 1910 Stanton (p. 188) discussed the relations of the Fox Hills sandstone to the Lance in eastern Wyoming and adjacent States and concluded:

"The bearing which the facts here presented have on the Laramie problem is self-evident. If it is true that there is a transition with practically continuous sedimentation from the Fox Hills sandstone into the Lance formation in the region discussed, then the Lance formation includes or forms part of the Laramie."

Finally Stanton (1914, p. 354) concluded a discussion of the Cretaceous-Tertiary boundary as follows:

"If, then, the Lance flora is in fact a Cretaceous flora, notwithstanding its close relationship with Eocene floras, it is obvious that the correlation of other formations with known Eocene formations on the evidence of fossil plants alone is open to serious question. In the case of the Denver [that is, the dinosaur-bearing portion] and Arapahoe formations such a correlation is directly opposed by the evidence of the vertebrate fauna, which allies them closely with the Lance formation."

Dobbin and Reeside (1930, p. 25), after a comprehensive study of the contact of the Fox Hills and Lance showing that alleged unconformable relations between these formations are either nonexistent or of a minor nature, conclude that:

"The Laramie and Arapahoe formations and at least part of the Denver formation of the Denver Basin, most of the Medicine Bow formation of southern Wyoming, and the 'Laramie' formation of northwestern Colorado, conformable on beds of Fox Hills age, are equivalent to the Lance formation."

Subsequent evidence does not contradict these statements. On the contrary, further comparisons of the faunas and floras (Dorf, 1942, p. 123) accentuate the similarity of the Laramie (enlarged) and Lance and confirm their time equivalence.

LARAMIE AND HELL CREEK

The type section of the Hell Creek formation is on Hell Creek, a tributary of the Missouri River, north of Jordan, Montana. The name has since been given to similar deposits in southeastern Montana and adjacent territory. The writer has examined many sections of these strata and has found that stratigraphically and paleontologically they are the equivalents of the Lance in eastern Wyoming. The Laramie, just shown to be equivalent to the Lance, thus becomes the equivalent of the Hell Creek also.

LARAMIE AND MEDICINE BOW

The geology and paleontology of the Medicine Bow formation have been treated recently by Dorf, who concluded (1938, p. 31, 37):

> "It is at once apparent from table 4 that the Laramie flora of the Denver Basin, Colorado, shows the closest relationship to the lower Medicine Bow flora, with 27 species in common.
>
> "The age of the lower Medicine Bow formation is rather definitely established by its stratigraphic position, its relation to the *Sphenodiscus* zone and to *Triceratops*-bearing beds, and its associated fresh-water molluscs. The Laramie and lower Lance (Hell Creek member) floras are shown to be most nearly similar to that of the lower Medicine Bow and are considered essentially contemporaneous."

Dorf used the term "lower" for the Medicine Bow formation as generally understood. Although he did not discuss in detail the relation of the Medicine Bow to the overlying Ferris formation, his southern Wyoming column in the correlation chart (1938, p. 36, Fig. 8) indicates that he did not consider the "lower" Medicine Bow to be the top of the Upper Cretaceous in that area but that there is presumably an "upper" Medicine Bow represented by the unnamed block of strata cut off from the basal Ferris. Veatch (1907, p. 527) asserted that an unconformity representing the removal of approximately 20,000 feet of strata separates the units now known as the Medicine Bow and Ferris formations. An unconformity of this magnitude at the designated stratigraphic level was considered by Knowlton and others as confirming their belief in a great diastrophic break between the Mesozoic and Cenozoic in the Rocky Mountain region. However, as in the case of the Laramie-Arapahoe unconformity, this alleged break has not been substantiated. Bowen (1917, p. 232) suggested that Veatch had inadvertently, by 6000 feet or more, mislocated the great unconformity, which is now recognized as that separating the Ferris and Hanna formations and representing erosion that beveled the exposed edges of more than 20,000 feet of inclined strata. Thus the supposed great unconformity between the Medicine Bow and Ferris is now regarded as nonexistent or as marked by a transitional zone indicating erosional irregularity. Moreover, the basal 1000 feet of the Ferris formation is a coal-barren conglomeratic zone whose

matrix and pebbles are similar to those of the Medicine Bow. This zone yields remains of *Triceratops* and a Cretaceous flora not materially different from that of the "lower" Medicine Bow and Lance. The writer has found specimens of the typical Paleocene flora in the lowest coal-bearing strata just above the latest dinosaur remains of the Ferris, indicating that the Cretaceous-Tertiary boundary is at or near that level. It is concluded, therefore, that the Medicine Bow formation encompasses the basal dinosaur-bearing portion of the Ferris formation and that the remainder of the Ferris, about 5500 feet, is Paleocene. Some basal strata of the Medicine Bow may be of Montana age.

LARAMIE AND VERMEJO

The Canon City coal field (Washburne, 1910) is an isolated Cretaceous-Tertiary area along the east front of the Rockies midway between the Denver Basin and the Raton Mesa coal field (Lee and Knowlton, 1917). The critical physical feature here and in the Raton Mesa field is a supposed unconformity between the Vermejo formation and the overlying beds. The writer confirms Washburne's description of the unconformable relations between the coal measures and overlying conglomerate in Alkali Gap (p. 348-350) and finds little or no difference between the stratigraphic position and lithology of the basal conglomerate here and the Arapahoe conglomerate of the Denver Basin. The little patch of Denver(?) overlying the conglomerate (Washburne, 1910, p. 350) contains only comminuted, unidentifiable plant fragments, and its age therefore remains doubtful.

In the Raton Mesa field Lee recognized an erosional boundary between the Vermejo and Raton formations, which was magnified into a considerable unconformity (1917, p. 55-56). The writer in 1939 verified the presence of the boundary and concluded that it is the time equivalent of that in the Canon City field and the post-Laramie pre-Arapahoe discordance on the western side of the Denver Basin. Diastrophic movements of the mountain area to the west were thus apparently simultaneously but unequally reflected in erosional activity along the Rocky Mountain front from Raton, New Mexico, to northern Colorado. This, however, was only an episode in Upper Cretaceous history and did not terminate the period.

Lee, depending upon Knowlton's study of the floras, held that the Vermejo formation is of Montana and the Raton formation of Tertiary age (1917, p. 56). The earliest strata now assigned to the Vermejo may be of Montana age, but the relationship of the Vermejo to the Laramie (enlarged) flora indicates that most of the Vermejo is of Laramie age. Knowlton based his conclusion that the Vermejo flora is Montana in

age on his identification of more Montana species than of any other age in the Vermejo. Of his 106 species in the Vermejo and 129 species in the Laramie of the Denver Basin he found only 6 species in common (1917, p. 231). The writer has restudied Knowlton's material in the light of new collections and has a tentative list of 29 species common to the two floras. This list may be enlarged after some undescribed species have been further studied. The original Vermejo flora reduces to 71 and the Laramie to 83 species. As the deposits in which the Vermejo flora occurs are only 40 miles south of the Laramie deposits on the southern edge of the Denver Basin and are stratigraphically at or near the same level, the large percentage of species common to the two floras suggests approximate contemporaneity. Accidents of preservation, the difference in latitude, and the relative positions of the floral localities in the formations, may account for inequalities in the composition of the floras. Thus the difference in latitude, provided zonal conditions prevailed in the past as they do today, may explain why the Hell Creek flora of Montana differs notably from the Vermejo flora of New Mexico and Colorado although they are time equivalents and have many "interfingering" species in common with the intervening contemporaneous Laramie and Lance floras.

Immediately above the Vermejo-Raton boundary are several hundred feet of relatively barren deposits, but above this zone are the productive coal measures of the Raton formation from which the bulk of the Raton flora was described. No paleobotanist questions that this is a Tertiary flora. Interest, however, centers on the basal, comparatively barren strata. Because no animal fossils and only a few fragmentary plants have been collected from this zone its age assignment is conjectural. One significant collection—U. S. G. S. no. 5794, made at the Bowen mine, locality 110 (Lee and Knowlton, 1917, p. 130, 278), a few feet above the basal conglomerate of the Raton formation—contains *Paleoaster inquirenda* Knowlton, a problematic plant referred to before as occurring high in the Laramie (enlarged) on Sand Creek east of Denver, and in the dinosaur-bearing part of the Dawson arkose near Ramah. As this fossil also occurs in the Vermejo but seems never to have been collected from horizons proved to be Tertiary, its presence in the basal Raton above the unconformity suggests that an indefinite thickness of the barren zone is of Upper Cretaceous age and should be included in the Vermejo formation.

SUMMARY

(1) The basal 290 feet of the Denver formation and 500 feet of the Dawson arkose contain dinosaur remains and Cretaceous plants. The beds immediately above these strata contain Paleocene mammals and

Tertiary plants and fall within the Puercan division. No mammals have been reported from the lower strata, and no dinosaurs from the higher strata.

(2) The lithologic boundary between the faunal units is generally recognizable, where exposed, by a change in color and is as satisfactory as most boundaries in a conformable continental series. It is better than the long-accepted boundary between the Arapahoe and Denver formations, for the first appearance of andesitic debris, said to mark this boundary, occurs at widely divergent levels in different localities. Neither a color nor a paleontologic change is evident at the Arapahoe-Denver boundary.

(3) The Laramie formation, as originally defined, has in recent years yielded, in addition to the well-known Cretaceous flora and nonmarine invertebrates, dinosaurs of the same types as those of the Arapahoe and Denver formations and the Dawson arkose. The Denver has recently yielded invertebrates hitherto found only in the Laramie in the Denver Basin. The Laramie therefore forms a paleontologic unit with the Arapahoe formation and the basal parts of the Denver formation and the Dawson arkose.

(4) The writer accepts the boundary between the Laramie and the Arapahoe and between the Laramie and the Dawson arkose as an erosional discordance or a transitional zone, depending upon the exposure observed.

(5) The writer proposes that the Laramie include the Arapahoe formation and the basal part of the Denver formation and the basal part of the Dawson arkose. It is believed that the continuity of the fauna and flora outweighs in importance the possible discontinuity in sedimentation.

(6) This Laramie formation as enlarged is equivalent to the typical Lance formation and the Hell Creek formation; to the Medicine Bow formation plus the basal portion of the Ferris formation; and to the Vermejo formation plus the basal portion of the Raton formation, with the exception that both the Medicine Bow and Vermejo as now defined may include in their basal parts some strata that should be referred to the Fox Hills sandstone.

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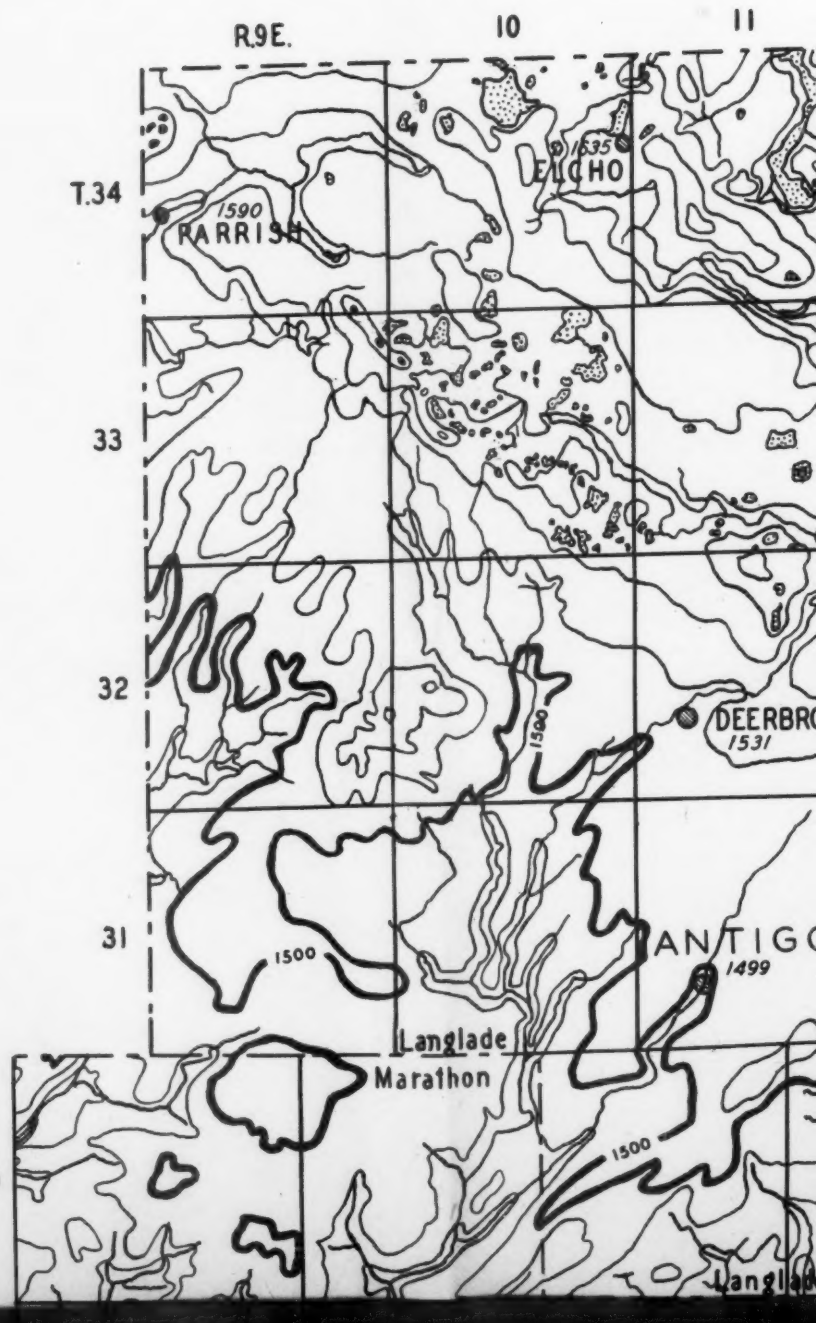
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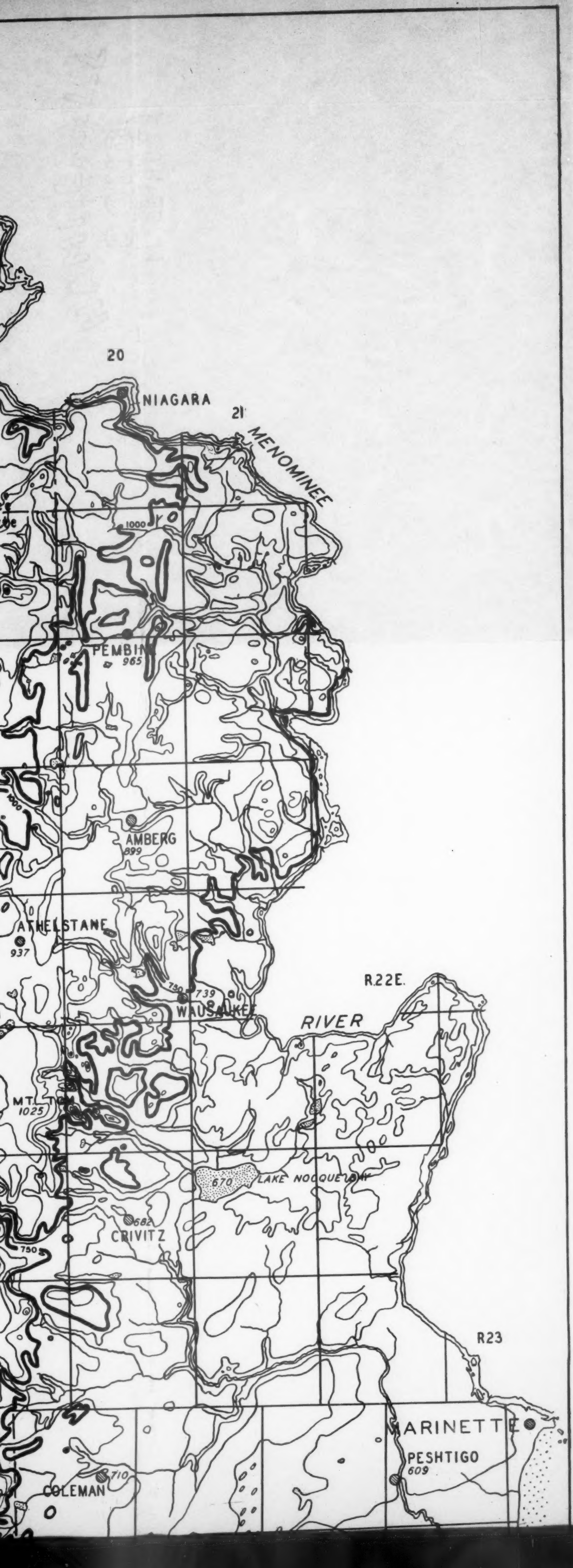
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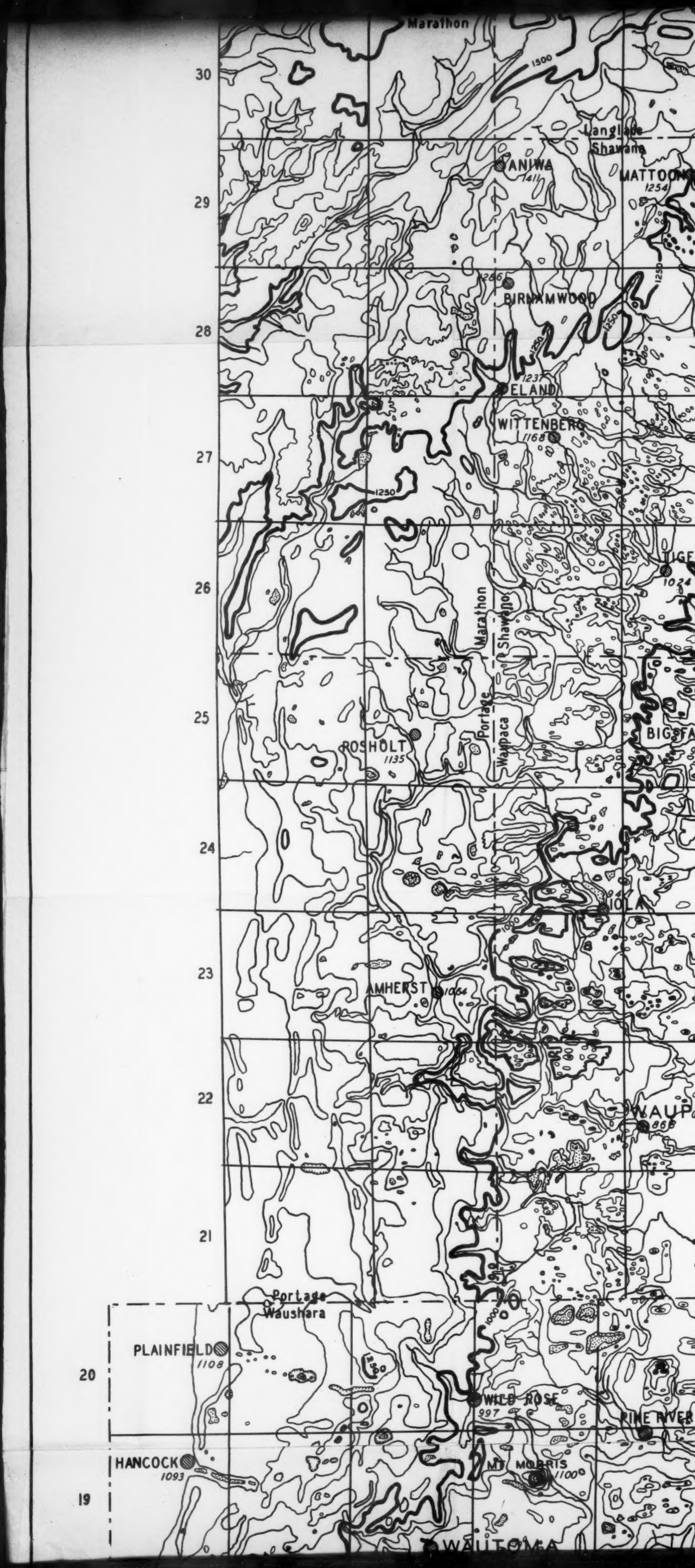
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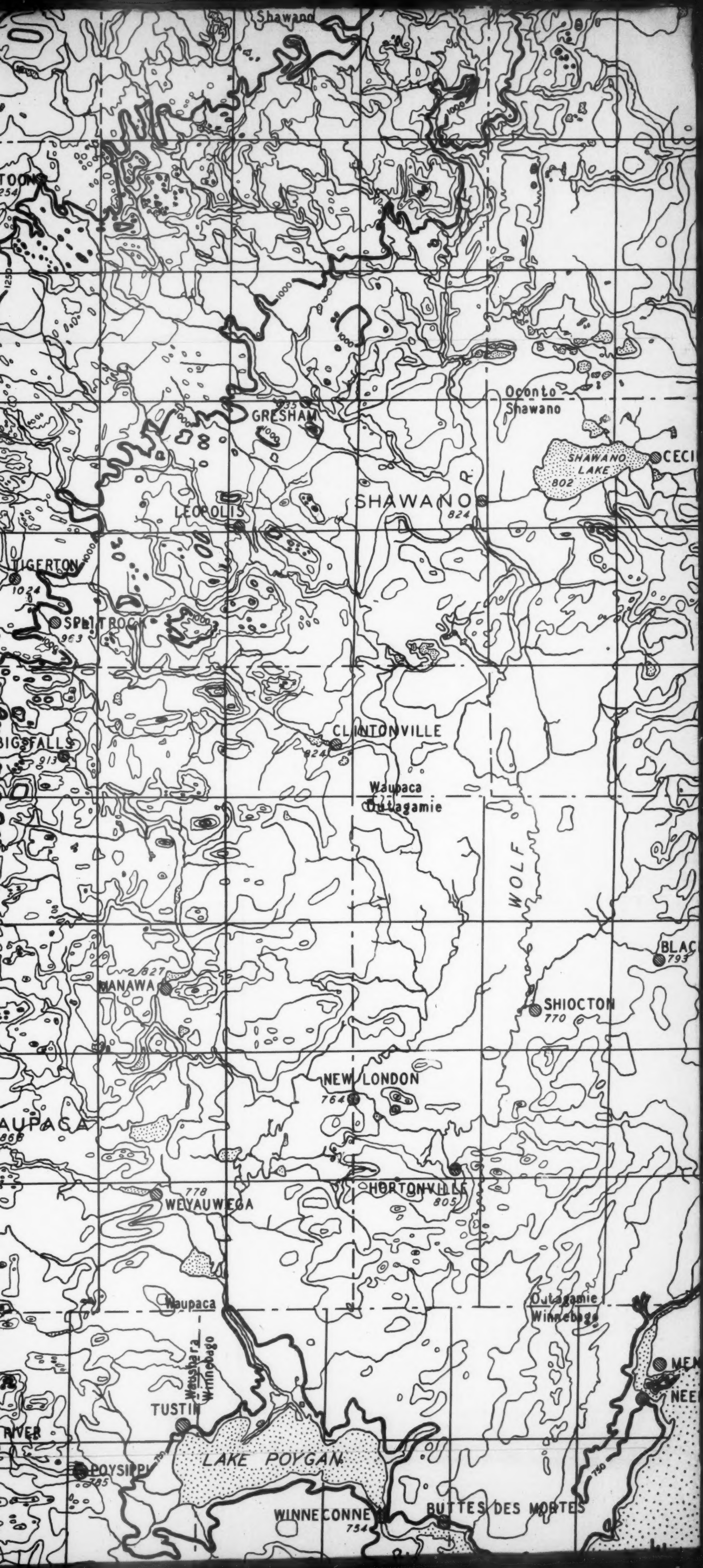
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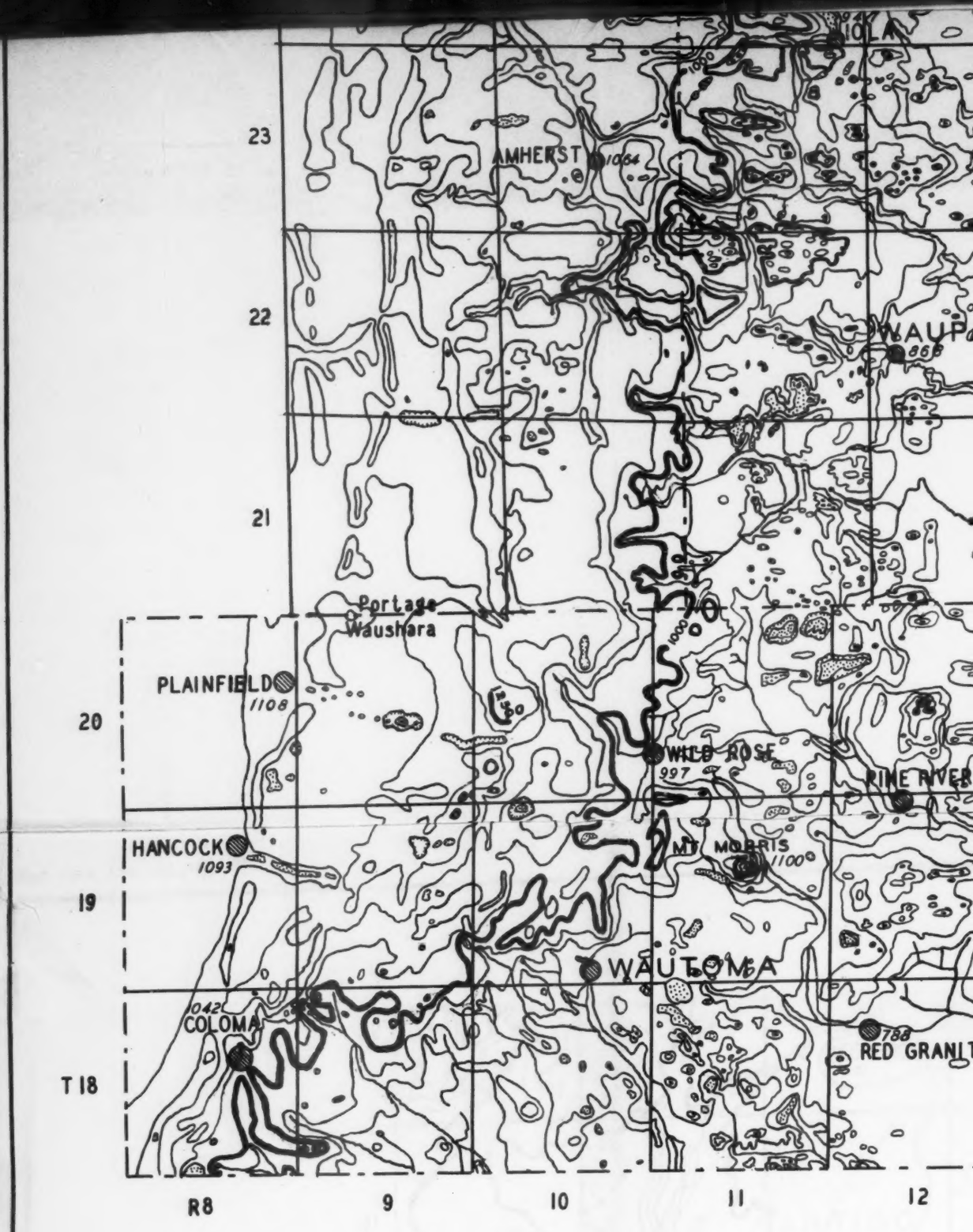
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F. T. THWAITES 1926-1941

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PLEISTOCENE OF PART OF NORTHEASTERN WISCONSIN

BY F. T. THWAITES

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ABSTRACT

The portion of northeastern Wisconsin discussed comprises that portion of the Green Bay lobe which lies north of the district described by Alden (1918), west of Lake Winnebago, Fox River, and Green Bay, and a portion of the Langlade lobe to the northwest. Outside of the drift which is bordered by the terminal moraine, long recognized as of Wisconsin age, is a thin, patchy drift almost devoid of glacial topography. Although this drift was long called "old" or "pre-Wisconsin" it is now thought that most, if not all, the extra-morainic deposits are Iowan, although some may be Illinoian.

The Wisconsin drift is divisible into deposits of the Cary and Valders substages. The name Valders is proposed in this paper because of the uncertainty of former correlations across the brush- and forest-covered region around Lake Superior.

The survey was a detailed reconnaissance. Every practicable effort was made to use both the criteria of topography and of material in the mapping. Aerial photographs were used throughout much of the area. Attention was devoted to the geochemical relations of the drifts to ground, stream, and lake waters. Some attention was also directed to the mechanical composition of the glacial and aqueo-glacial deposits. The history of glacial dissipation is illustrated with a series of block diagram maps. The glacial lake of the Fox-Wolf valley is renamed Glacial Lake Oshkosh.

HISTORY OF INVESTIGATION

Examination of the Pleistocene deposits of part of northeastern Wisconsin (Fig. 1) was begun by the writer in 1926 as part of the program of the Wisconsin Geological Survey for the development of local materials for highway construction. State support was withdrawn after 3 field seasons. A little work was done in 1931 with the aid of students from the University of Wisconsin. Field work from 1934 to 1940 was financed by a grant from the Penrose Bequest of The Geological Society of America, thanks for which are here given. Assistants in field and office work were Loyal Durand, II, C. G. Dickinson, R. J. Koplin, G. T. Owen, A. T. Eberhardt, V. U. Hanson, P. W. Icke, P. J. O'Neil, C. A.

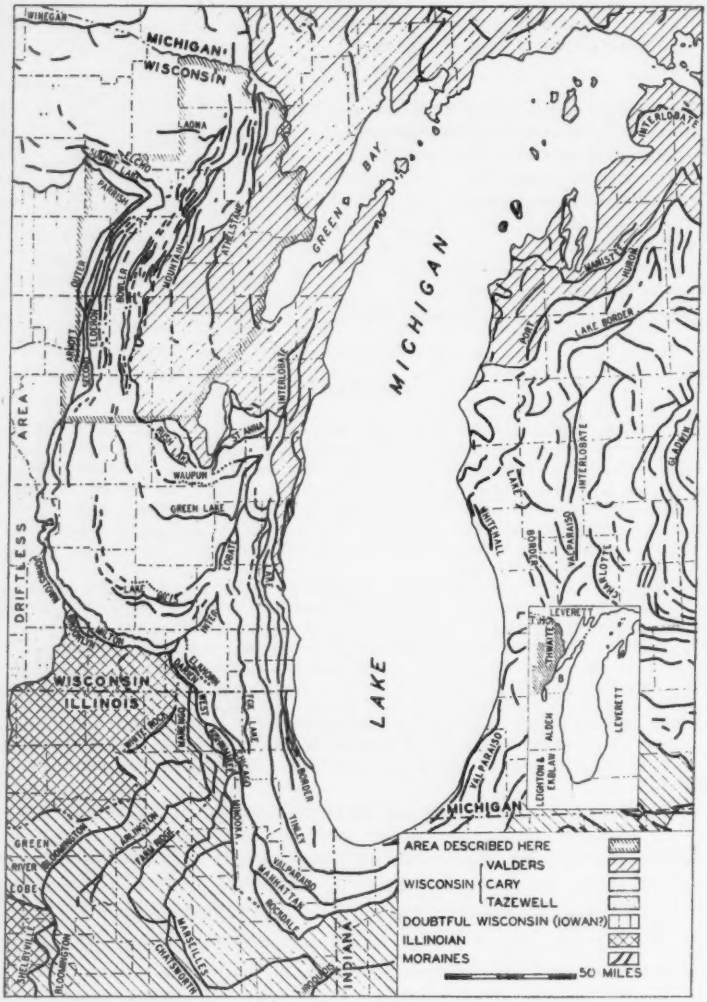


FIGURE 1.—*Location of area and moraines around Lake Michigan*

Based on published surveys by Alden, Leighton and Ekblaw, Leverett, and unpublished work of Kenneth Bertrand and A. J. Hanners.

Bays, Herbert Von Eiff, Edwin Wenberg, Kenneth Bertrand, and Amy M. Thwaites. Stuart Carter did some work near McCaslin Mountain in 1940, the results of which have been incorporated in Plate 1.

FIELD METHODS

GENERAL STATEMENT

The present survey is classed as "detailed reconnaissance" intended to work out the geological history without taking time to complete exhaustive studies and detailed mapping with the aid of numerous test pits and borings. Owing to the fact that the region examined varies from open, treeless plains to dense brush or "green timber," the amount of travel per unit of area which was necessary to obtain the desired results varied widely from place to place. Transportation was mainy by automobile along roads and trails, although in some localities considerable foot traversing was needed along abandoned logging roads, logging railroads, and through brush. During the earlier years of the survey many miles, not only of trails but in some places of well-traveled highways, had to be traversed with the compass to place them on the map. Since 1934, however, the surveys of the State Conservation Commission, the United States Forest Service, and the Civilian Conservation Corps provided invaluable data to supplement the very unsatisfactory published maps of the little-settled portions of the area. Since 1939 the aerial photographs of the United States Geological Survey and the Agricultural Adjustment Administration were available at the office of the State Highway Commission. Intensive study of these stereoscopic views, coupled with visits to some of the new fire towers, cleared up many of the problems of the earlier part of the work. Complete aerial checking was done in Florence, Forest, Marinette, Oconto, and Langlade counties, but time did not permit extension of this work to all the area.

ELEVATIONS

Only a small part of the district here discussed has been mapped topographically by the United States Geological Survey (Wausau Special and Neenah quadrangles and a narrow strip of Ripon and Neshkoro quadrangles), and much of that was very crude work done long ago. Planimetric maps without contours based on aerial photographs are now in preparation but were not completed in time to revise the base map for this report. The shores of Peshtigo River and Menominee River were contoured by the State Geological Survey and the United States Engineer Corps, respectively (Smith, 1906; 1908; Herron, 1917; Skinner, 1922; Trippe, 1932). The shore line of Green Bay and some of the adjacent

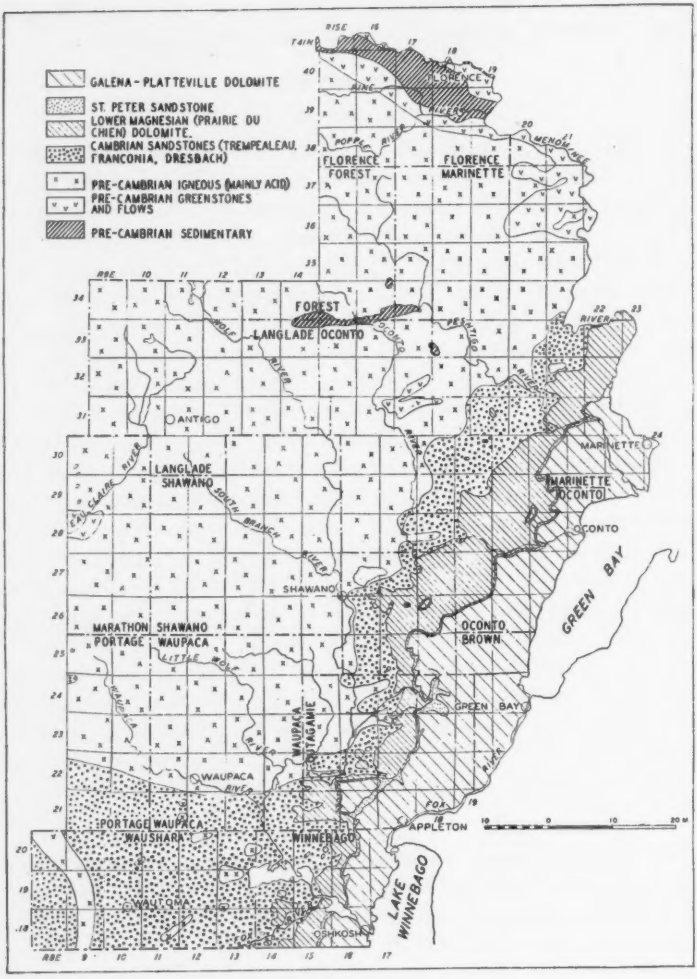


FIGURE 2.—*Bedrock geology*

Part of northeastern Wisconsin. Age relations of different types of pre-Cambrian rocks not given. Based on information available in files of Wisconsin Geological Survey, 1940.

land was mapped by the United States Lake Survey. Profiles of railroads were secured from the chief engineers of the several operating companies to whom thanks are due. A few bench marks of the United States Coast and Geodetic Survey are found in the area. Using these data for control, elevations were extended throughout the district by aneroid using Lahee's (1920; 1941) method of adjustment. Most of the observations and computations were made by the assistants. The accuracy of this work varied greatly both with weather conditions and with the "personal equation" of care and thought devoted to the problem. It is believed that few elevations are more than 20 feet in error. The topographic map (Pl. 1) was based on elevation data supplemented by notes on slopes made in the field. It does not pretend to show all details, and only a small part of it was corrected by the use of aerial photographs.

PREVIOUS SURVEYS

Much of the area here described was never reached by the earlier geologists, Irving, Chamberlin, and Brooks. The western part of the area was resurveyed by Samuel Weidman between 1897 and 1906 (Weidman, 1903; 1907). Work by William C. Alden on the region farther south overlapped a portion of the south edge of the area now under discussion, but many of his data were not included in his published map (Alden, 1918). Copies of his notes and field maps have been examined in the office of the State Geological Survey in Madison. In 1910 W. O. Hotchkiss resurveyed Florence County, and his field notes are on file. The soils have been mapped throughout the area, mainly by the State Geological Survey in co-operation with the United States Bureau of Soils (Whitson, 1916-1929). The soil maps were the best base maps available at the time of most of the field work, and Plates 1 and 10 are based on them with minor corrections from other sources. Goldthwait (1907) examined some of the abandoned shore lines west of Green Bay. In 1933 H. R. Aldrich, then Assistant State Geologist, completed the survey of Langlade County begun by the writer. He kindly allowed the use of his data in order to complete the maps herewith. Road-material surveys under the direction of E. F. Bean, State Geologist, have been drawn upon for screen tests of gravel and other data. Other reports which bear on the district are those on clays by Buckley (1901), on limestones and marls by Steidtmann (1924), and on underground waters by Weidman and Schultz (1915) and by the State Board of Health (1935).

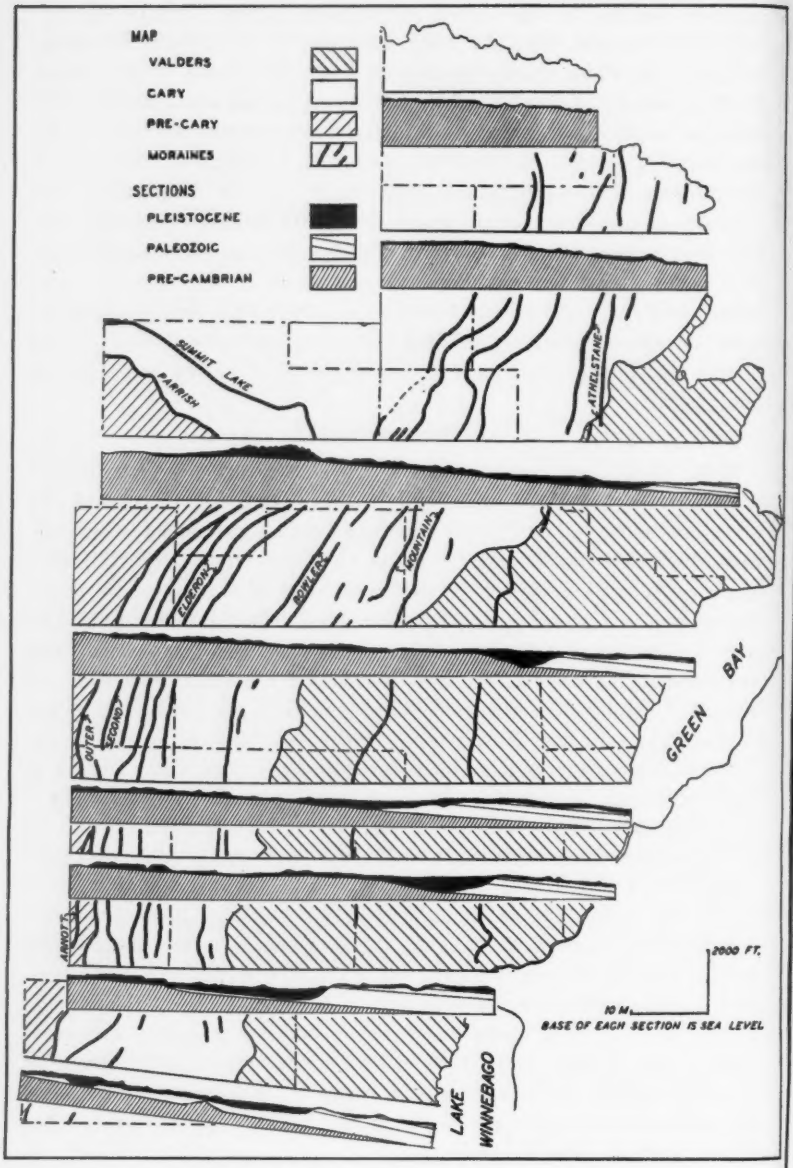


FIGURE 3.—*Cross sections showing structure*

BED ROCKS

INTRODUCTION

More than half the area here described is underlain by pre-Cambrian bedrock (Fig. 2). The remainder has Upper Cambrian to middle Ordovician sedimentary rocks. Exposures are generally rather poor, although there are some wide expanses of bare rock and very thin drift (Pl. 5, fig. 1). Good sections of the sedimentary rocks are rare. Recourse has been had to samples taken during the drilling of water wells, most of which are in the sedimentary area.

PRE-CAMBRIAN

Granites of many types and probably several ages are by far the most abundant pre-Cambrian rocks. Many are very coarse-grained; pegmatitic and aplitic dikes are abundant. The greenstones are altered lavas and are older than most, if not all, of the granites. Hornblende schist, gabbro, and various porphyries occur in small quantity. Pre-Cambrian metamorphics are present as small areas in Florence County and the high bluffs locally called McCaslin and Thunder (Blue) mountains. These rocks consist of slate, tuff, quartzite, and iron formation of Huronian age. Iron ore was formerly mined in Florence County, and the other Huronian areas have been explored for it. Scarcity of iron formation in the drift strongly suggests that, if any concealed ranges are present, they must be small. Some of the granite at least intrudes the sediments (Van Hise and Leith, 1911; Leith *et al.*, 1935). The map changes in the border of the pre-Cambrian are based in part on well records and in part on the eastern limit of boulders which are different from those in the drift where it lies on Paleozoic sedimentary rocks. Drift boulders are also much more abundant in the crystalline district. These criteria suggest that many outliers of the sediments on the rolling surface of the pre-Cambrian have been omitted in Figure 2.

PALEOZOIC SEDIMENTS

The sedimentary rocks of the Paleozoic rest unconformably on the peneplaned surface of the pre-Cambrian (Thwaites, 1931) and dip gently south of east (Fig. 3). The following column is based mainly on subsurface study (Thwaites, 1923; Twenhofel, *et al.*, 1935):

ORDOVICIAN	Thickness Feet
Galena—dolomite, light gray where fresh, buff on weathered surface.....	100
Platteville (including Decorah, also called Black River in some reports)—dolomite, blue gray in two horizons, remainder light gray where fresh, buff on weathered surface, sandy at base.....	100

ORDOVICIAN—Continued	Thickness (feet)
St. Peter—sandstone, fine- to medium-grained, mainly white, surface yellow-brown to gray, thickness very irregular	0–125
Lower Magnesian (Prairie du Chien)—dolomite, light gray, some red layers, much sand and green shale; chert, both oölitic and dense; bedding and thickness very irregular with numerous mounds, probably reefs; locally absent	0–283
CAMBRIAN	
Trempealeau—dolomite, mainly red, sandy and silty; locally a few feet of clean medium-grained gray sandstone at top (Jordan member)	33– 66
Franconia (formerly Mazomanie)—sandstone, mainly fine-grained, bottom coarse-grained light gray, dolomitic, in part glauconitic	80–141
Dresbach (Galesville and Eau Claire members)—sandstone, medium- to coarse-grained, a few layers of shale	0–265

In Figure 2 the border of the Lower Magnesian was mapped on the basis of well records plus physiography, and an error in previous maps along Menominee River was corrected.

PLEISTOCENE

METHOD OF APPROACH

The Pleistocene deposits of northeastern Wisconsin may be approached as: (a) land forms, and (b) sediments. It is apparent from most literature on glacial deposits that land forms were predominantly used in the past. As a matter of fact, it is the only possible approach in regions of few exposures and in the study of aerial photographs (Thwaites, 1934, p. 27-32). A complete study of the problem of genesis must also include data on chemical composition, bedding, and sorting. In fact, certain kinds of deposits can be identified only by their composition and not by their topographic form. Knowledge of unconsolidated deposits is hampered by (1) the ephemeral nature of exposures so that new excavation is needed to obtain accurate data, and (2) the time and effort needed to make analyses from the necessarily large samples. The writer made no mechanical analyses in the field except pebble counts (Fig. 4) but obtained his data from those of the road-material survey. As the purpose of the latter survey was economic, only those materials had been tested which appeared to be potentially usable. Starting with a 50-pound sample only 1 pound of the material which passed a ¼-inch screen was tested. Although till has been used for surfacing some very sandy roads in this area, no screen tests of it have been made. Some data were obtained from soils analyses (Whitson, 1927; Kellogg, 1930), but most of these show only the A and B horizons, and there is always some doubt where the C horizon is named that it actually included only

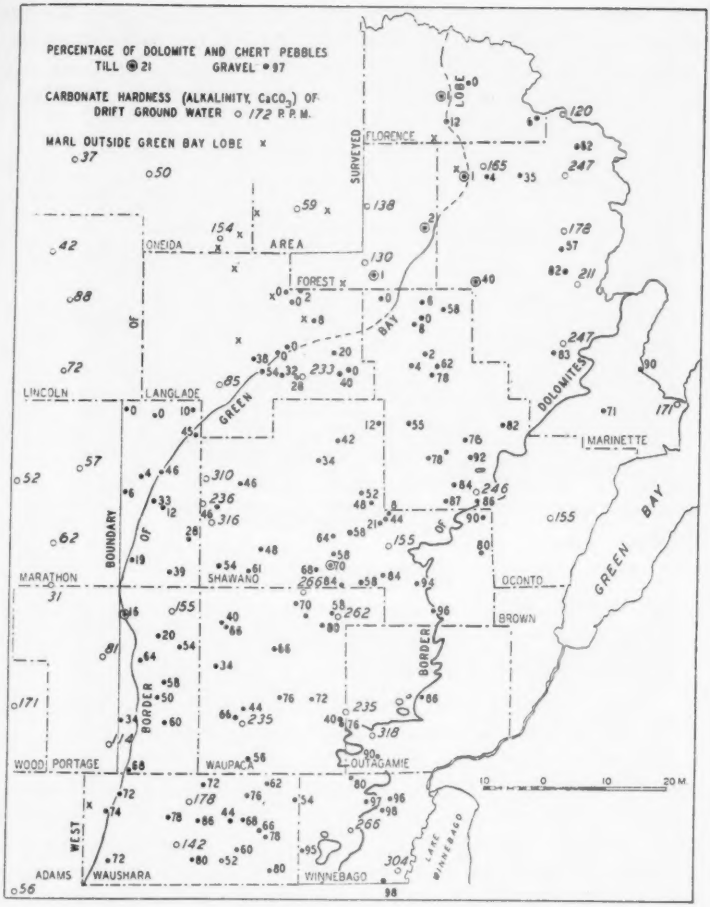


FIGURE 4.—*Relation between dolomite content of drift and hardness of drift ground waters*

Data on waters from Wisconsin State Board of Health and Weidman and Schultz.

unaltered material. Soil analyses were based on small samples and show only the finer materials. Alden (1918, p. 322-323) gives a few mechanical and chemical analyses of clays and tills. Buckley (1901, p. 270-274) gives many chemical analyses of clays (Fig. 5). Pebble and boulder counts do not give an accurate picture of the drift because the nature

of the parent material from which they were derived controls the size distribution of fragments in the drift more than do the various processes of transportation and sorting. For instance, most granites break into large boulders because joints are far apart, although some coarse-grained

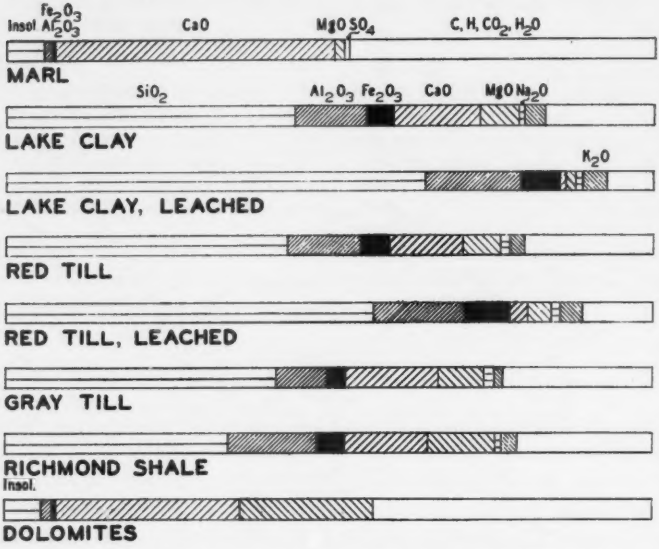


FIGURE 5.—*Chemical composition of marls, clays, till, and bedrocks*

Based upon analyses published by Steidtmann, Buckley, and Alden.

varieties furnish on disintegration much material in the granule range. On the other hand, limestones or dolomites commonly form pebbles from ¼ inch up to a few inches in diameter. Large boulders of limestone are rare. Most of the sandstone of northeastern Wisconsin is so soft that it appears only in the drift as sand which will pass a 20-mesh screen.

MAPPING

Practical field mapping of unconsolidated deposits is difficult over much of northeastern Wisconsin for the reasons given above. If undue weight has been given by the writer to surface forms, it was simply because of the scarcity of exposures throughout much of the district and because of the lack of time for digging test pits through the weathered zone which is several feet thick. In brush and standing timber ground views of the topography are so limited that errors in interpretation are inevitable.

FIGURE 1. UNPITTED OUTWASH OF ANTIGO FLATS AND VERY ROUGH TERMINAL MORAINE OF GREEN BAY LOBE

Near Bryant, Langlade County, at corner of T.32 N., R.11 E., T.32 N., R.12 E., T.31 N., R.11 E., and T.31 N., R.12 E. Wooded areas on outwash occupy a fan of stony gravel deposited outside gap in sec. 8.

FIGURE 2. RIDGES IN TERMINAL MORAINE

Northeast of Lakewood, Oconto County T.33 N., R.16 E., and T.33 N., R.17 E. These ridges are not eskers; they parallel the ice margin and are till. They are marginal deposits possibly accentuated by shove due to advance of ice margin. Lake in sec. 13 is in pitted outwash. Curved swamp in secs. 6 and 8 is a drainage line.

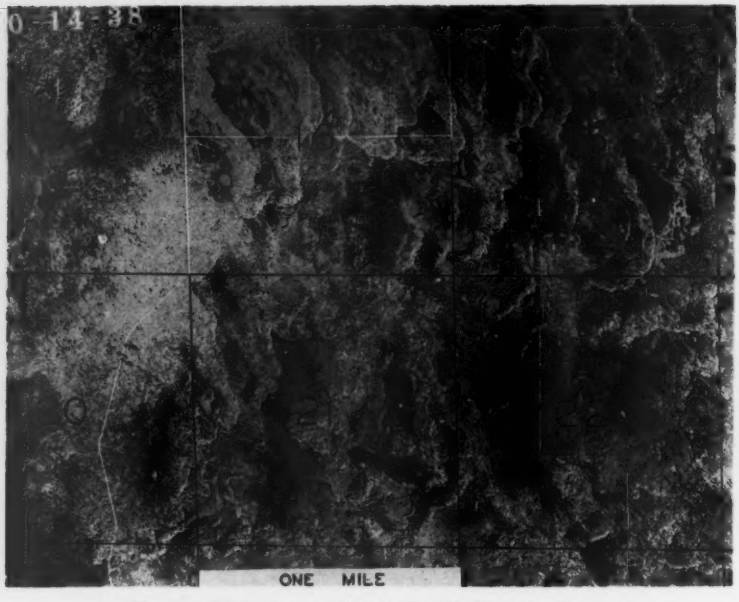


FIGURE 1. TOPOGRAPHY IN T.38 N., R.18 E., FLORENCE COUNTY
Brush-covered land in secs. 16 and 21 is a terminal moraine with many N-S ridges. Hill in south part sec. 19 is ground moraine probably with rock core. Remainder is pitted outwash with rounded kettles and sandy soil.

FIGURE 2. TERMINAL MORAINE
Green Bay lobe of Cary substage. Southwest of Lakewood, Oconto County (SW1/4NW1/4 sec. 7, T. 32 N., R.16 E).

TOPOGRAPHIC FEATURES IN WISCONSIN

Those who may in the future check the mapping by excavation or drilling will undoubtedly find many discrepancies due to the fact that the material shown on the map is very thin at that particular point. An attempt was made, however, to map the materials which are buried beneath the relatively thin Valders red till.

LAND FORMS

General considerations.—Land forms of the Pleistocene and Recent in northeastern Wisconsin are here classified into terminal moraine, ground moraine, drumlins, outwash, eskers, lake-shore features, lake plains, stream terraces, and sand dunes. Although most of these may be recognized by their diagnostic shapes, some require a knowledge of the internal character and sedimentary environment for proper determination.

Terminal moraine.—Terminal moraines in the area consist of ridges up to several hundred feet high whose topography is for the most part extremely chaotic and irregular. Knobs, kettles, and minor ridges, in many cases enclosing depressions, occur in what at first seems a haphazard arrangement. Views from the air show, however, that the arrangement is not entirely without plan (Pl. 2; Pl. 3, fig. 1) as it seems from the ground (Pl. 3, fig. 2). The minor ridges trend roughly parallel to the ice front. Their development is irregular in degree, and some moraines show only confused knolls with little trace of a linear pattern. Prominent ridges are common in the vicinity of Lakewood in Oconto County (Pl. 2, fig. 2). Such ridges appear to be till which was banked against the ice margin; these may be due to shove by a minor readvance. The most abundant material in moraines is till. In the crystalline region the till is very bouldery (Pl. 4, fig. 2). Locally coarse bouldery gravels occur in moraines. Where the ice margin was in standing water, as near Green Bay, deltas accumulated (Pl. 7, fig. 1). The moraines are subject to many interruptions due to burial by outwash, erosion by glacial meltwater, burial by later ice advance, or simply due to nondeposition because the ice front did not maintain its stand along a definite line long enough. Each moraine appears to represent a rejuvenation of the ice sheet with consequent readvance of the margin. In many deep excavations outwash or lake clay is shown under the till of a moraine. In one case in Oconto Falls varved clay had been disturbed by ice whose till had been eroded before terrace gravels concealed the record. The smooth outline of the ice fronts recorded by most moraines also indicates readvance, for a very irregular border would certainly result from melting back. Some areas mapped as terminal moraine which are not linear may be errors due to inclusion of areas of overridden pitted outwash. Viewed

in cross section (Fig. 3) the moraines of the Green Bay Lobe form a gigantic staircase whose treads are outwash plains and whose risers are moraines. In many places the crest of the till in the moraine is lower than the surface of the outwash deposited beyond the ice.

ARNOTT MORAINE: The Arnott moraine (Fig. 1, Pl. 10) was discovered by Weidman (1907, p. 456-459, 605-606) who correctly interpreted it as much older than the Outer moraine a short distance to the east. It is a ridge about 50 feet high and a mile wide which occurs in several sections. Scattered pebbles are found on top of Mosquito Mound near Bancroft just west of the map (Weidman, 1907, p. 558-562). The till was derived from the east, for it contains stones from the Paleozoic bedrocks and the very coarse-grained granites of that area. Dolomite pebbles have been destroyed by weathering to a depth of over 10 feet, and this weathered zone is slightly cemented by iron oxide. The till was formerly used for road surfacing. Wind-cut stones have been reported by Thiesmeyer and Digman (1942).

Ground moraine.—The term "ground moraine" as here employed represents: (1) Places where little drift was deposited (Pl. 5, fig. 1); (2) spots from which meltwaters eroded most of the drift; (3) rock-controlled hills with a moderate drift cover; and (4) areas of till overlying older glacial and glacio-fluvial deposits. Very little of the present ground-moraine surface consists of deposits made under the ice, but most is till laid down when the ice melted. The topography is in general inherited from underlying deposits. Figure 1 of Plate 4 shows a terminal moraine buried under later ground moraine which was not thick enough to conceal the original form. Figure 2 of Plate 5 shows typical ground moraine in the district underlain by red Valders till, and Figure 2 of Plate 6 shows ground moraine on older outwash. Part of Figure 1 of Plate 3 is a high ground moraine which is probably a rock hill.

GROUND MORAINE OUTSIDE THE TERMINAL MORAINE: Outside the terminal moraines of the Green Bay and Langlade lobes the drift is relatively thin and overlies crystalline rocks. On the divides the surface is nearly level, and large areas are marshy. (*See* Wausau special topographic sheet.) Some low constructional knolls occur in western Langlade County, but over most of the area they are wanting. The total relief is several hundred feet and increases westward toward the Wisconsin River. Narrow gravel terraces border the streams. In the past such topography has almost universally been considered erosional. As there is no sharp break between uplands and valleys, it seems more reasonable to conclude that the topography is simply a reflection of the pre-Cambrian peneplain, dissection of which increases toward the west.

Drumlins.—Because the primary diagnostic feature of drumlins is form, they are difficult to determine in brush- and forest-covered country. Even within the well-settled parts of northeastern Wisconsin most of the drumlins are wooded. This accounts for the fact that, although several

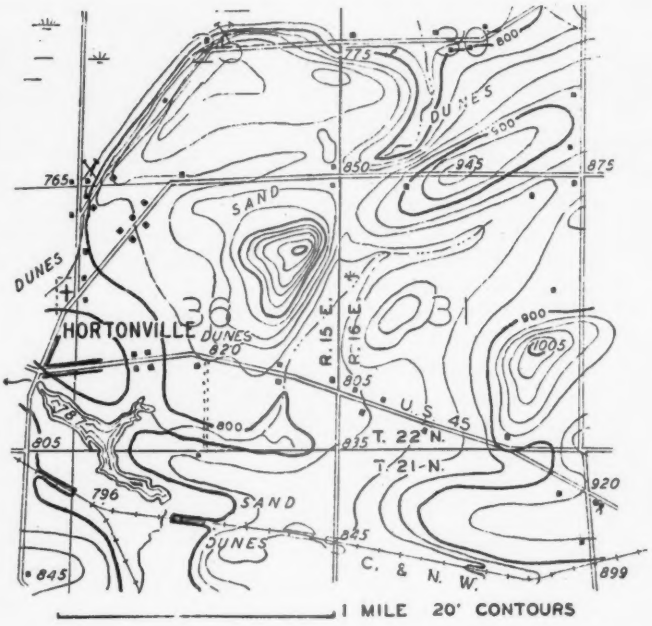


FIGURE 6.—*Overridden drumlins near Hortonville, Wisconsin*

Drawn from aerial photographs, aneroid observations, and railroad surveys. Southwestward-moving Valders ice built new tails on Cary drumlins.

geologists had visited the region previous to the writer, none of them recognized the drumlins. Examination of aerial photographs confirmed the ground surveys of the writer. Almost all the drumlins are composed of till, although a few were discovered where there is a substratum of glacio-fluvial deposits. A good example of this phenomenon is found along the County Highway south of Marion, Waupaca County (NW ¼ sec. 14, T. 25 N., R. 13 E.) The sides of many drumlins have been scarred by postglacial gullies. In other places glacial and glacio-fluvial deposits which formed as the drumlins melted through the ice have combined to obscure the form. In many places several adjacent drumlins are more

or less fused so that "drumlin uplands" result. Individual members of such a group are best distinguished from a distance where the crests may be seen. Even in the aerial photographs separation is not readily made. Rock-cored hills may resemble drumlins, but such are believed to be rare in this region. Perfect drumlins are very abundant and range up to 250 feet high and over 2 miles long. From this maximum there is a downward gradation in size and perfection to faint flutings which cannot be mapped. Within the area of Valders till few drumlins have much relief, and all are difficult to see on the ground. Many of those shown on Plate 10 were distinguished only in the aerial photographs. Discrimination of just which hills to map as drumlins is a matter of personal judgment, and no two geologists would reach exactly the same conclusions.

Overridden Drumlins: Within the area covered by the Valders red till there are many drumlins of unusual form ascribed by the writer to a change in direction of ice movement (Fig. 6). The cores are gray till, evidently deposited during the Cary substage when the ice moved almost due west. The ice which deposited the red till came from the north or northeast and made new tails on many of the drumlins. In some places, for instance southeast of Shawano, the older drumlins were largely obliterated and new ones built of red till. To understand how ice passes over unconsolidated deposits without removing them it is necessary to realize that glacial erosion is effected by plucking and not by abrasion (Matthes, p. 50-102; Thwaites, 1934, p. 19-21). The till within a drumlin is so well packed and free of joints that it could only be removed en masse which was apparently beyond the power of the ice. For this reason the overriding ice altered the drumlins mainly by addition.

MATERIALS OF TERMINAL, GROUND MORAINE, AND DRUMLINS

General statement.—The land forms designated as terminal moraine, ground moraine, and drumlins are primarily composed of till. The writer confines the term till to debris which was deposited directly by the ice without rearrangement by water. Although till is unstratified, it does not follow that all drift which lacks bedding is till. Sorted deposits may never have been stratified because of continuous deposition or may have had layers mixed by slumping or weathering. The glacial tills of northeastern Wisconsin vary widely in physical constitution as shown in Figure 7. The silt-clay component varies widely. It is at a maximum in the red till which closely resembles lake clay in mechanical composition. Unfortunately no satisfactory analyses are available of till in the crystalline region, but apparently the finer materials are in considerable proportion although large boulders are conspicuous (Pl. 4, fig. 2). In the area of

FIGURE 1. CARY TERMINAL MORAINE
Covered with Valders red till, northwest of Seymour, Outagamie County (SE1/4 SW1/4 sec. 3, T.24 N., R.17 E).

FIGURE 2. VERY BOULDERY TILL IN REGION OF CRYSTALLINE ROCKS
Near Polonia, eastern Portage County (SW1/4SE1/4 sec. 5, T.24 N., R.9 E).

TERMINAL MORAINE AND TILL

FIGURE 1. GREENSTONE LEDGES IN GROUND MORAINE
T.37 N., R.21 E., Marinette County. Topography is controlled by rock structure and not by ice flow. Pitted outwash south of outcrop area.

FIGURE 2. ESKERS IN T.31 N., R.21 E.
Near Loomis, Marinette County. Shadows are on north sides of ridges, and some persons will find it necessary to turn this plate upside down to see the relief.

GROUND MORAINE AND ESKERS

sedimentary bedrock the silt-clay is at a minimum where the underlying rock is sandstone. Comparison with Figure 5 which shows chemical composition demonstrates that the gray till is more like the Richmond shale than like the dolomite, although Figure 4 shows that virtually all the pebbles come from the near-by dolomites. On the sandstones the per-

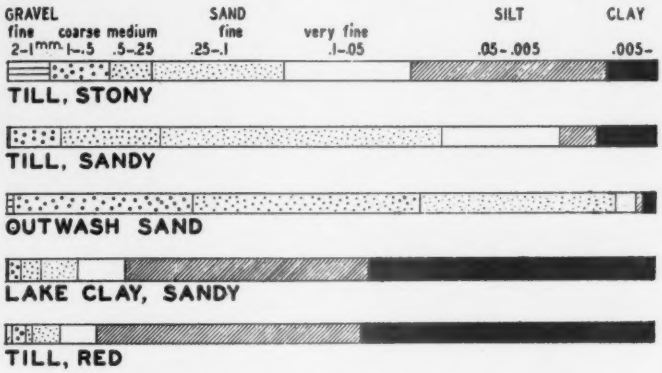


FIGURE 7.—*Mechanical composition of tills, sand, and lake clays*
Similarity of red till to lake clay is evident. (*Compare* with Figure 5.)

centage of dolomite pebbles does not decrease very markedly with distance from the source. Where the bedrock is granite, however, more local contribution to the pebbles is evident. Even the most sandy tills contain enough silt and clay so that excavations in them stand for a number of years with a vertical face.

Soils on till.—Soils developed by surficial alteration and weathering of till vary widely within the area. Published definitions do not in all cases separate soils on till from those derived from sorted materials. In general, however, the more sandy members of each series were not derived from till. On the extramorainic drift the soil has very impervious B and C horizons which are both mottled light gray and yellow brown. Such soils are called Colby or Spencer (Whitson, 1929; Kellogg, 1930, p. 67-71). The amount of chemical alteration is slight, however, and these soils are not gumbotil although subsoil drainage is poor. They are classed as *planosol*. Apparently the original high clay-silt content of the till plus the occurrence of impervious bedrock at relatively shallow depths caused the difference between these soils and the lighter Kennan soils on and within the terminal moraine (Kellogg, 1930, p. 83-85). The data of Figures 4 and 8 show that much more of the till is calcareous at depth

than was formerly supposed, although the till of the Green Bay lobe has by far the larger amount of material derived from dolomite. The source of the carbonate rocks of the Langlade lobe to the northwest is uncertain. Possibilities are: (1) dolomite near Limestone Mountain,

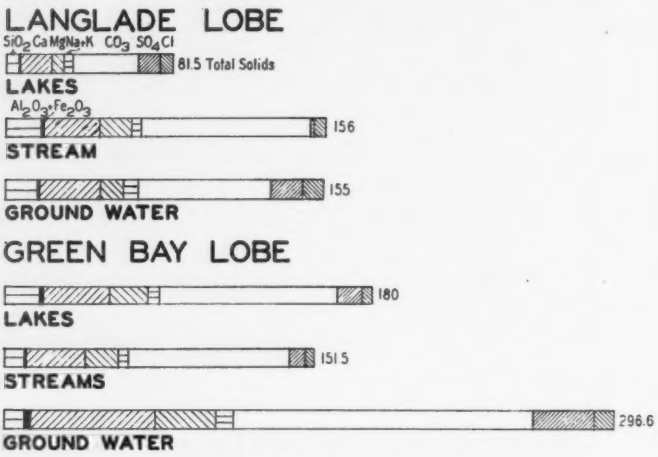


FIGURE 8.—*Mineralization of lake, stream, and ground waters* In Langlade and Green Bay lobes. Data from Weidman and Schultz.

Michigan, (2) calcareous Keweenawan shale in the Lake Superior basin, (3) dolomite and limestones in the Winnipeg region far to the northwest, and (4) the dolomites and limestones of the northern peninsula of Michigan to the northeast. Soils derived from the calcareous till of the Green Bay lobe are mapped as Miami or Bellefontaine, the latter on rougher land. Both are confined to a small area in Winnebago County (Kellogg, 1930, p. 37-41, 43-46). Farther west where the till is very sandy lies the type locality of the Coloma soils. The soil of the red till or reworked lake clay is included on some maps with soil on undisturbed lake clays as Superior; on other maps the till soil is discriminated as Kewaunee (Kellogg, 1930, p. 75-83). The old map of northeastern Wisconsin erroneously called much of the red till soil which has a deep A horizon "Miami" or "Coloma" (Kellogg, 1930, p. 73-74). The soils of northeastern Wisconsin are transitional between the gray-brown forest group and the podsols. The most evidence of podsolization (Kellogg, 1930, p. 19-21), which involves bleaching of the "A" horizon with iron oxide cementation

below, is found northwest of Green Bay where the forest cover was densest.

Outwash.—Outside the end moraines of the Cary substage of the Wisconsin drift the outwash (Pl. 1) forms an almost level plain. Where streams debouched from the ice front leaving a break in the moraine a low fan was deposited. Material in such fans contains more stone than the average of the plain, and one of the fans may be distinguished in the aerial view (Pl. 2, fig. 1) by the fact that it was left in woods whereas the rest of the outwash is cultivated. Inside the end moraines, however, deposition was over and around many ice masses which varied widely in size and abundance and were residual rather than transported in the shallow outwash streams. Melting of the included ice blocks led to a complex of knobs, mesas, kettles, and ridges or crevasse fillings (Pl. 6, fig. 1; Pl. 7, fig. 2). Some ice blocks were concealed by outwash and left rounded shallow depressions with neither boulders nor till at the surface. Others projected and thus protected considerable areas from outwash deposition. Meltwaters, in part from the ice masses, eroded the deposits and added to the complexity. In many places terracing took place before all the included ice had disappeared so that the several levels are all pitted. The topography of pitted outwash is in many places much like that of terminal moraine, and discrimination must be made on the basis of material. The pitted outwash knolls and mesas all show fairly well sorted, horizontally bedded outwash. However, samples of the pre-outwash topography exist in the bottoms of kettles, some of them nearly a mile across, or when meltwaters eroded a large share of the deposit, as near Tigerton, then the perplexity of the geologist is extreme. Should he attempt to separate the protected areas, many of them till or bedrock, or generalize as outwash? In other places so many rock ledges or till hills project through an outwash plain that they can hardly be mapped separately. In this case should the generalization be as ground moraine or as outwash? Decision to generalize many areas as outwash was based upon the economic purpose of the work, on the lack of detailed maps when field work was in progress, and on the small scale of the final map. Had the aerial photographs been available in the field, a much finer discrimination could have been made. The possibility of confusion of pitted outwash with terminal moraine is serious first, in brush covered regions, and second, where the outwash was overridden by a readvance of the ice (Pl. 6, fig. 2). In the latter case exact discrimination would require so much subsurface data that where the till is too thick to be penetrated in road cuts a final decision proved impracticable. Another source of confusion consists of ground moraine

hills or drumlins whose tops rise only slightly above the surface of the outwash. For these reasons Plate 10 does not check with adjacent maps by Weidman (1907) and Alden (1918). Doubtless this confusion of pitted outwash with terminal moraine explains many of the errors of soil classification in the older reports.

Subglacial wash.—VerWiebe (1927) who worked in Menominee County, Michigan, immediately north of the area here described, suggested that much of the assorted drift of that region which shows no definite plane surface accumulated under the glacier. A similar theory was also advanced by Andersen (1931; *see also* Thwaites, 1934, p. 47). Their evidence appears to have been (1) the large number of boulders which they assumed to have dropped from overlying ice, and (2) the scarcity of definite proof of slump due to melting of buried ice. After years of careful search the writer is not convinced that such a thing as "subglacial wash" actually exists, at least in northeastern Wisconsin. The possibility of such large subglacial cavities is questionable, and so many cuts show outwash sands in which bedding has been disturbed or destroyed by slump that the second argument seems to have little basis. Boulders are normally present in outwash as noted below.

NATURE OF GLACIO-FLUVIAL DEPOSITS

General statement.—In order to understand glacio-fluvial sedimentation it is necessary to review some of the fundamental conditions. It is obvious that glaciers normally gave off floods of water especially during decay and that these waters found available a large amount of loose material. It is also evident that the amount of water derived from ice melting varied widely not only from day to night but from winter to summer. Some glacial streams flowed in ice-bound channels, either tunnels or open cracks, whereas others could spread out over flat surfaces. Some of the deposits of these rapidly varying, turbulent, and commonly fully loaded streams have since been reworked by waves, postglacial streams, or the wind. Figure 9 gives a graphical comparison of several types of sorted drift with a till. Water transportation of drift reduced the silt-clay content because it was carried most readily and therefore was transported farthest from its source or to a quiet environment where it could settle. Removal of the finer components of the till left the esker and kame gravels relatively high in stones over 2 inches in diameter. Bedding in such deposits is confused, and the degree of sorting varies widely among different beds. Most eskers of the district have much stony gravel, although northwest of Green Bay certain eskers are very sandy.

Glacial meltwater deposits on land outside the ice.—The term outwash is herein confined to the deposits of glacial meltwaters on land outside

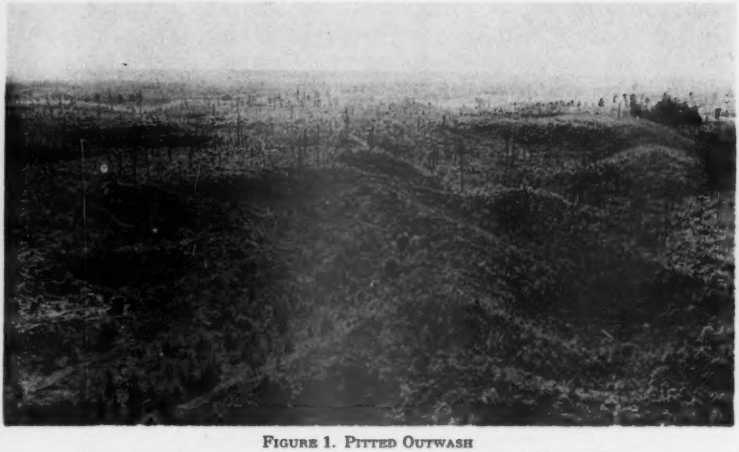


FIGURE 1. PITTED OUTWASH
Showing fillings of cracks between stagnant ice masses. From top of Newald Firetower, east of Newald, Florence County (SW1/4NE1/4 sec. 36, T 38 N., R.15 E).

FIGURE 2. PITTED OUTWASH
Covered by Valders red till. Northwest of Oconto Falls, Oconto County (SW1/4SE1/4 sec. 10, T.28 N., R.19 E).

PITTED OUTWASH

FIGURE 1. DELTA OF EARLY LAKE OSHKOSH
Covered by red Valders till, southeast of Seymour, Outagamie County.

FIGURE 2. OUTWASH, APPARENTLY OF VALDERS AGE
Near Wausaukee, Marinette County, in T.34 N., R.20 E., and T.34 N., R.21 E. Many crevasse fillings deposited between closely spaced ice blocks.

the ice margin. Two different environments of deposition occurred: (1) on land which had not been previously occupied by ice or had not been occupied for a considerable time; and (2) on territory incompletely freed of such ice. To have deposition by streams the velocity must have been less than that necessary to transport the coarser particles.

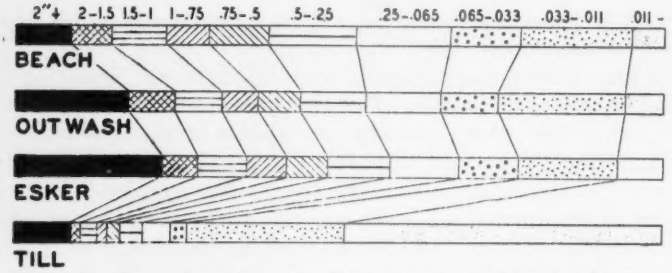


FIGURE 9.—*Comparison of gravels of different origins with till at Madison, Wisconsin*

Data from Wisconsin Geological Survey files.

This discussion does not consider ice rafting which, although glacial streams were shallow, unquestionably accounts for many large stones which are out of harmony with the remainder of the deposit. Not all glacial drainage resulted in deposition immediately on leaving the ice margin, nor did all glacier-fed streams deposit continuously along their courses away from the ice. A good example of such discontinuous deposition is along Eau Claire River. The extensive Antigo Flats of outwash were laid down above the narrow, constricted section of the river at the Dells. Below that section of high velocity, deposition was resumed (Pl. 10). Weidman (1907, p. 523, 525) apparently did not recognize the occurrence of outwash above the Dells, for he states: ". . . the Eau Claire river . . . has alluvial terraces only along its lower course, terminating some distance below the Wisconsin moraine. . . ." The grade at which outwash was deposited in the region here discussed varies from a maximum of about 40 feet per mile in some of the alluvial fans of coarse material near the ice margin down to only about 4½ feet per mile along Wisconsin River between Wausau and Mosinee west of the area described. For the first 10 miles beyond the ice the average grade is about 12 feet per mile. It is difficult to measure accurate grades on many outwash deposits because contemporaneous stream positions are hard to correlate. However, the figures show that the dip of the bedding is too low to recognize in a single exposure. Cross-bedding is common

especially in sandy layers and for the most part dips in the average direction of stream flow. The streams which laid down the deposit were of the braided type causing a wide variation in direction of dip of foresets. Cross-bedding rarely extends through more than a few feet of strata. Distortion of bedding by slump due to melting of ice masses buried in the outwash is common. The degree of sorting may be termed "fair" (Fig. 9) and is notably better than in eskers. It is not at all difficult to discriminate typical outwash gravel by either bedding or by screen test. Very commonly outwash gravels have the highest pebble content at or near the surface. This phenomenon has been explained in several ways: (1) most gravel pits were started at places where stony gravel reached the surface naturally; (2) gravel pits are located on the crests of bars which underwent concentration by subsiding glacial streams which removed the sand; and (3) wind removed some of the sand from the outwash. Deposits coarser than the average of a given plain occur in narrow ridges which were deposited by streams confined between ice masses. Adjacent to kettles where ice blocks projected above the adjacent filling the outwash was locally reworked and the pebbles concentrated by meltwaters from the ice blocks. Ice masses also contributed boulders to the deposit. For the reasons advanced above the presence of boulders does not prove that till underlies the surface at that point. In many cases excavation has demonstrated that the boulders were originally scattered through the outwash and were concentrated by the later erosion of the finer parts of the deposit. Outwash deposits of the age of the extramorainic drift are rare.

Soils on outwash.—Most of the sand in northeastern Wisconsin contains many other minerals than quartz, and therefore weathering formed soils having considerably more silt and clay than the parent material. Soils on outwash vary from the very sandy Plainfield series, where possibly wind removed much of the silt and clay as fast as it formed or the original content of quartz was uncommonly high, through the very rough Vilas series, to the much better Fox, Waukesha (prairie), and Antigo soils (Whitson, 1927; Kellogg, 1930, p. 46-49, 56-57, 61-66, 87-90). Many of the areas mapped on the older surveys as the more sandy members of the Coloma and Kennan series are now known to be soils derived from outwash and not from till. The confusion appears to have arisen because borings were not carried through the weathered zone, because of ice-rafted boulders or boulders derived from buried ice masses, and through a misunderstanding of the topography as explained above. In general, the outwash soils are far more subject than till soils to crop failure in dry years because of the loose texture of the underlying mate-

rial. In many places soil erosion has made serious inroads in areas of sandy soil because of the friable subsoil.

Eskers.—In this report the term "esker" is applied to any ridge of assorted materials which is definite and reasonably continuous. Short

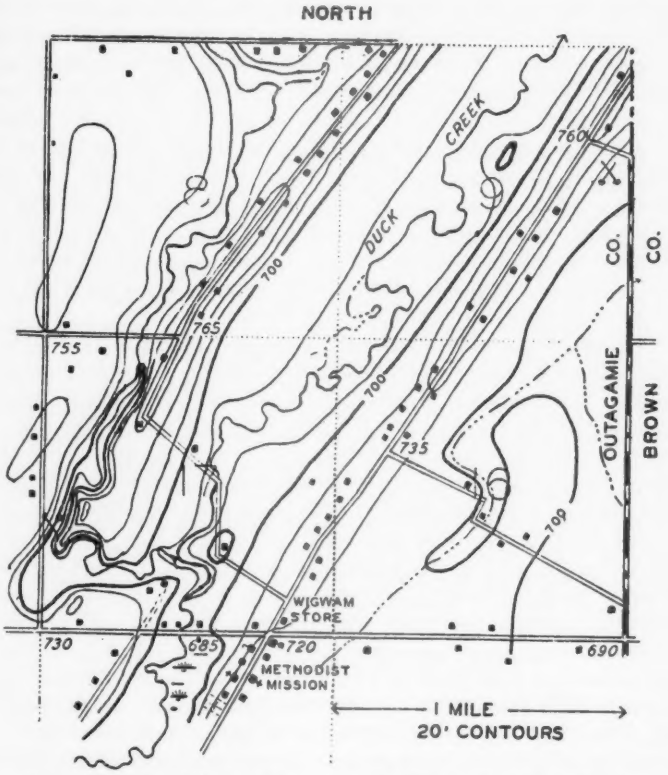


FIGURE 10.—*Overridden eskers or crevasse fillings along Duck Creek*

T. 23 N., R. 19 E. Map drawn from aerial photographs controlled by aneroid elevations.

ridges of sand and gravel within a pitted plain (Pl. 7, fig. 2) which obviously represent fillings between closely spaced ice blocks have been ignored in Plate 10. Some of the ridges in outwash areas which were mapped from the aerial photographs may be of this class but were shown on account of their length and prominence. Eskers shown on Plate 10 attain a maximum length of nearly 20 miles but are not continuous, for

they are divisible into several sections. Few eskers are as much as 50 feet above the adjacent till. Many meander as shown in Figure 2 of Plate 5. Exceptions to this are the long, straight, twin eskers which enclose the valley of Duck Creek southwest of Green Bay (Fig. 10). These were formed during the disappearance of the Cary ice and overridden by the Valders readvance. Subdivision, branching, and rejoining of eskers, in many places at right angles to the general trend, are common phenomena, but it is impracticable to show all such on the scale of Plate 10. Although some eskers rest on a surface which declines toward the west (Pl. 5, fig. 2), many rise notably toward the western terminus, and a few cross over small drumlins. Yet the streams which formed them certainly flowed west as proved by transportation of stones and dip of foreset bedding. The ends of few eskers have any connection with moraines or other marginal accumulations. The Duck Creek eskers grade into deltaic moraine at their southwest ends. Most eskers which were wide enough to escape the effect of slump of the sides show a top which is essentially level over considerable distances.

Environment of Esker Deposition: That eskers represent the deposits of glacial meltwaters which flowed between ice walls from which much of the debris was derived is self-evident. The position of these streams with reference to the ice sheet and the state of that ice are, however, matters of dispute. In the area here discussed the bedding of eskers has been disturbed only by slump along the sides proving that the material was deposited where it now lies and not on top of or within the glacier. Furthermore, the dendritic form of many of the eskers (Pl. 5, fig. 2) certainly indicates that the adjacent ice was motionless. Whether the channels were exposed or roofed by ice is less readily decided. Many of the short eskers are very irregular in height and broken into many sections, suggesting deposition in a tunnel. Many of the long eskers lie on land which rises toward the west. The esker which extends from the Marinette-Oconto county line just north of Lena rises at its base about 150 feet in 16½ miles. It is broken into four distinct segments each of which has a level top. It may very well have been deposited in separate sections by a stream in an open crack. If so, there is no distinct evidence of progressive lateral escape of waters at the several breaks. Possibly the ice border was so inconstant in position that no trace of such escape should be expected. Many eskerlike deposits, not all of them notably coarser than the material of adjacent plains, are what Flint termed crevasse fillings. They blend imperceptibly into adjacent outwash. A plexus of such forms is shown in Figure 2 of Plate 7. Others, generally much coarser and stonier, are true eskers partially buried by outwash after the melting of the confining walls. Several

eskers occur in the very rough overridden moraine of eastern Shawano County. This fact demonstrates that the rough topography is inherited from an older Cary moraine (Pl. 4, fig. 1), for it is difficult to imagine esker deposition under the conditions which prevailed just at the edge of the ice during the formation of the knolls. Surely an environment which formed a confused mass of small hills would not permit formation of reasonably continuous esker ridges. A few ridges of till which resemble eskers were discovered and are regarded as filling of crevasses where there was not enough water to sort the material derived from the adjacent ice.

Lake shores.—In very few places within the area are abandoned lake shores a conspicuous feature of the landscape. Only here and there, mainly on islands or peninsulas where wave action was particularly effective, can beach ridges and wave-eroded cliffs be distinguished with certainty. It seems probable that Lake Oshkosh was so much obstructed by ice that wave action was very local in its effects. Even the shores of Lake Algonquin and Lake Nipissing are weakly developed along the west side of Green Bay (Goldthwait, 1907, p. 95-98). This lack of development is due partly to shallow water close to the shore and partly to offshore winds. The Algonquin shore has undergone tilting and rises from approximately elevation 610 at Green Bay to about 630 west of Marinette. This distance is about 50 miles so the slope is not over 6 inches to the mile. The Nipissing shore is so weakly developed just below elevation 600 that it is difficult to determine its exact location. Postglacial beaches of Shawano Lake are 6 to 12 feet above lake level and contain abundant shells. Frank C. Baker (personal communication, 1926) listed from this beach in the SW¼NE¼ sec. 17, T. 27 N., R. 17 E.: *Naiad* fragments; *Campeloma*, immature, aff. *rufum* (Haldeman); *Stagnicola* cf. *Catascopium* (Say); *Helisoma antrosa* (Baker).

Lake plains.—Although shore features are hard to see in much of the area, there is a marked topographic contrast between the flat lake bottoms and the more rolling glacial topography which was never submerged by standing water. Off the mouths of large streams like the Peshtigo and Menominee, however, stream features due to meandering are impressed on the lake bottom topography. These abandoned channels are readily observed in the aerial photographs although difficult to observe from the ground. Lake basins occur in the abandoned lake beds of which Lake Winnebago, Lake Poygan, and Shawano Lake are some of the best known examples. All these lakes are very shallow. Typical kettles do not occur in the lake beds. Where no kettles are found in a sandy plain and stones are small or absent, it is in many

places hard to decide between outwash, whose surface has been reworked by wind, and a lake plain. Marsh deposits are not mapped for two reasons: (1) Their limits on some of the existing maps are very inaccurate, and it would have taken a prohibitive time to correct them, and (2) they would serve only to confuse the representation of the underlying drift.

Deposits of glacial meltwaters in standing water.—Wherever glacial meltwaters started to flow over an area newly or incompletely vacated by the glacier, they found an ungraded, irregular surface. The initial course of almost every stream was interrupted by areas of standing water. Melting of included ice masses formed many ponds and lakes across former stream courses. These facts should be kept in mind to evaluate the sporadic occurrence of both long foreset bedding and quiet-water sediments within areas which on the surface appear to be outwash. Extensive lakes are not proved by such phenomena. It must also be realized that many bodies of standing water were interrupted by residual ice masses for a considerable time after they first formed. All lakes close to the ice were undoubtedly frozen over for a large portion of the year and even in summer contained much drifting pack ice. Deposits in standing water may be divided into: (1) beaches and deltas deposited adjacent to the shore, and (2) deep-water sediments.

Deltas: Delta deposits may be distinguished from outwash by the long and deep foreset bedding (Pl. 7, fig. 1) due to the abrupt slackening of velocity of the streams where they reached standing water. Theoretically the sorting of the deposit should be decidedly less perfect than that of outwash, but screen tests to show this are not available for this area. Several of the deltas evidently were deposited around grounded bergs, for instance that east of Clintonville. On Plate 10 deltas deposited along the ice margin have been included with moraines and others with outwash.

Beach Deposits: Discrimination of the gravel deposits formed by waves from those deposited by streams is not easy because both came from the same parent materials. Study of Figure 9 shows, however, that beach gravels have much less coarse stony material and proportionately more below ¼ inch than do either outwash or kame gravels of the same region. In other words, sorting has gone further than in outwash because of a longer period of action by an agency which does not normally change so rapidly in power. The diurnal variation of glacial meltwater streams contrasts sharply with the long periods of moderate wave action between occasional storms. However, beach gravels have thinner layers than are common in outwash (Pl. 8, fig. 1). Many of the layers are almost

FIGURE 2. VARVED CLAY OF LATER LAKE OSHKOSH New London, Outagamie County. Light layers are summer and dark are winter components (NW1/4 NW1/4 sec. 6, T.22 N., R.15 E).

FIGURE 1. GRAVEL IN BAR OF LATER LAKE OSHKOSH Stephensville, Outagamie County. Dark layers are very well sorted "buckwheat" gravel (NW1/4SE1/4 sec. 17, T.22 N., R.16 E).

DEPOSITS OF LATER LAKE OSHKOSH

FIGURE 1 SAND DUNES ON OUTWASH PLAIN
Western Marinette County, T.35 N., R.18 E. Dunes look like waves and are of the blow-out type due to wind erosion where vegetation was locally destroyed, possibly by forest fires.

FIGURE 2. ICE CONTACT FACE
"High Plains" of Waushara County from Cemetery Hill, east of Wautoma, Waushara County (SE1/4NE1/4 sec. 35, T.19 N., R.10 E.).

DUNES AND ICE-CONTACT FACE

perfectly sorted "buckwheat" or "pea" gravel, but screen tests of single layers are not available. Other layers, for the most part thicker than these, are much less well sorted and probably represent deposits made during storms. They contrast sharply with the product of the constant lapping of small waves. There is, however, a notable scarcity of coarse, stony gravel in most beaches. Beach deposits of gravel are in few places more than 8 or 10 feet thick. In places beach sorting may be superimposed on gravels of other origin—because for a time they formed a shore—as for instance the top of the foresets in a delta. Beach deposits cannot be distinguished by absence of sand in a gravel bed constituting "open-work" gravel. Gravels from which sand was removed by increased stream velocity or in which it may never have been deposited occur also in eskers, outwash, and postglacial stream terraces. In such deposits the sorting is not so perfect as in real beach gravel. Many of the beach deposits of northeastern Wisconsin show a very high silt content due to the tenacious red clay of the till from which they were washed. In the road material tests the silt content is determined from wet examination of a 1-pound sample of the material which is smaller than ¼ inch, and the result does not show the silt which sticks to the larger pebbles. Occasional layers of clay interbedded with gravels are winter deposits. Bedding in a lake-shore deposit is horizontal in a cut parallel to the beach. The lake side of a bar has a gentle dip toward deep water. The lagoon side has foreset bedding in deltas formed by storms washing over the crest. In northeastern Wisconsin beach deposits occur mainly on islands and points where there was no floating ice to damp the waves.

Deep-water Deposits: Below the wave base only silt and clay were deposited. Shallow lakes, however, contained only sand. The cold meltwater direct from the ice apparently floated on top of the slightly warmer lake water, and winds descending from the ice blew this muddy water over all parts of the lake which were free of ice. The marked seasonal variation in supply combined with slow settling in cold fresh waters caused a very perfect size separation between summer and winter deposits. Summer layers are a gray or light pink silt (Pl. 8, fig. 2) with up to 25 per cent rock flour from the dolomites. Winter layers are fine red clay with much more iron oxide and up to 50 per cent dolomite flour (Ellsworth and Wilgus, p. 106-108). The red color of the finest material undoubtedly came from the red rocks of the Lake Superior region to the north.

Varve Succession: Partly because there are only a few good exposures of varved clay in northeastern Wisconsin, but mainly because the purpose of the survey when started was economic, only one varve section—that at Richford, Waushara County—was measured by the

writer. A section at Waupaca was measured by Ellsworth and Wilgus, and one at New London by Antevs. DeGeer attempted to place the New London section with respect to the Waupaca measurements, but geologic study definitely proves that this deposit is much younger than either of the others. The clay at New London was derived from the Valders till and differs markedly from that at Waupaca both in chemical and physical characteristics (Ellsworth and Wilgus, p. 110).

Local Absence of Lake Sediments: Throughout large portions of the areas mapped as "submerged by glacial lakes" no lake sediments could be discriminated. In some localities excavation would probably reveal deposits which were not apparent on the surface, but in other places it is clear that such is not the case. Areas without lake sediments may be accounted for in several ways: (1) In shallow parts of the basin wave action removed material as fast as it accumulated; (2) in other places a permanent cover of ice or a residual of the glacier prevented sedimentation; (3) currents in narrow straits prohibited deposition. In shallow parts of the lake discrimination of lake sand from outwash is difficult. A soil auger was used in some of the lake beds, but drilling through either dry sand or sand below water level is impracticable.

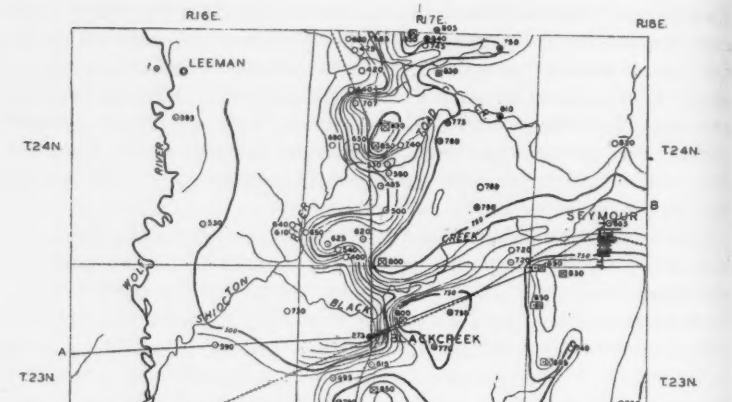
Soils on Lake Sediments: Soils which have developed on lake clays are mapped as Superior or Poygan where marshy. The sandy members of this series are due to a layer of sand above the clay (Cornucopia, Orienta). The soils on lake clays have been confused on many maps with those on the red till which is lake clay reworked by ice. Although published mechanical and chemical analyses (Figs. 5, 7) appear to show little effect of the ice on the parent material, washing a small sample almost universally shows more sand and small pebbles in the till than in the lake clay. This criterion is very useful in examining well samples and could have been used to separate the soils. The lake clays contain over 80 per cent clay but are not so impervious as the Colby soils which lie upon a till with a large content of silt as well as clay. The horizons of these lacustrine soils are almost the same as those on red till soils (Kewaunee) (Kellogg, 1930, p. 75-83, 90-94). Soils on deep lake sands where underlying clay, if present, is not readily discovered, have been mapped along with soils on outwash. In Outagamie County lake silt was erroneously mapped as Antigo.

Late glacial and postglacial sediments.—It is difficult in much of northeastern Wisconsin to draw a sharp line between Pleistocene and Recent deposits. Marls and peat are certainly postglacial. Marl is high in calcium and low in magnesium (Fig. 5), thus contrasting sharply with the glacial lake clays. Marl is of chemico-organic origin from material

in solution, whereas the dolomite rock flour of the glacio-lacustrine sediments was mechanically transported from the parent till. Marl is abundant throughout the Green Bay Lobe where the ground and stream waters are hard (Figs. 4 and 8) because of abundant dolomite debris in the drift even where bedrock contains little carbonate. It is not closely confined to that area, however, because there is some dolomite in the Langlade lobe area, and outwash from the Green Bay lobe extends far beyond the limits of the ice.

STREAM TERRACES: Many of the larger rivers of northeastern Wisconsin have terraces which slope downstream. Although the present positions of streams draining to Green Bay could not have been assumed throughout their entire length until the ice was completely gone, a gradual change took place from southerly or southwesterly drainage along the ice front to the steeper flow to the southeast. This is explained under the section on glacial history with a series of block diagrams. Buried ice masses survived in the outwash until after the upper parts of streams took the present courses. For this reason no hard and fast distinction between glacial and postglacial terracing is possible. Most of the later terraces show no kettles, and in many places the sorting of their deposits is better than that of ordinary outwash. With the exception of terraces and flood plains along the Oconto River and Duck Creek few postglacial terraces could be mapped with confidence. These show an eastward flow by their topographic position and direction of dip of cross-bedding. Some of the lower terraces along Brule, Menominee, Peshtigo, and other rivers cannot be correlated with assurance because the topographic position is such that glacial meltwaters might conceivably have contributed to the streams which laid them down. Gravels of apparent postglacial origin vary from very coarse to fine gravel which somewhat resembles that of a beach. Screen tests of gravels along Oconto River showed, however, no difference from glacial outwash; in other words the source of the material obscured the process of deposition. The better sorting of some layers is ascribed to the more even flow of present-day streams during low water when fed by lakes and springs compared with the highly variable flow of glacial meltwaters.

DUNES: In view of the abundance of sand in the glacio-fluvial and glacio-lacustrine deposits of the area under discussion, the presence of large dune areas is to be expected. In some dunes, however, the source of the sand is not readily ascertainable probably because the original deposit has been so extensively reworked or completely buried under dunes. Figure 1 of Plate 9 shows dunes where the sand was blown out of holes in the outwash due to local lack of vegetation. These breaks in vegetation may very likely have been due to forest fires since the



ROCK OUTCROPS
DOLOMITE
SANDSTONE

WELL RECORD
DRIFT
DOLOMITE
SANDSTONE
GRANITE

ELEVATION OF ROCK SURFACE OR BOTTOM OF WELL 800

50 FT. CONTOURS ON BED ROCK

FIGURE 11.—*Bedrock topography near Black Creek, Wisconsin*

timber was cut. Most of the dunes were anchored by vegetation when white men entered the region. In most localities the dune topography is complex, closely resembling that of some terminal moraines. In other places east-west ridges are common; they may be distinguished from eskers by the absence of pebbles. In one place in Outagamie County

crescentic dunes with convex side to the southwest suggest barchanes made by southerly winds. In general the distribution of dunes, which may be traced east on to rock hills or terminal moraines, combined with observation of the direction of dip of foreset bedding indicate westerly or southerly winds. Doubtless dune formation was spread over a long time. On Plate 10 the overprint does not give the exact limits of the dunes. In some districts, for instance north of Shawano Lake, the dunes have not entirely obliterated the pre-existing topography. Over much of the Valders drift the sand cover is in large part wind blown. Soils on dunes have been mapped as Plainfield, Vilas, and Coloma, the last two obviously errors.

PLEISTOCENE AND RECENT HISTORY

PREGLACIAL TOPOGRAPHY

General statement.—Several difficulties were met with in a study of the preglacial topography of northeastern Wisconsin: (1) Much of the northern part of the area is thinly settled, and there are few wells; (2) over large portions of the settled regions almost all the wells are either drive points or "tubular" wells of small diameter which end in drift; (3) many of the wells in the southern part of the area are so old that information about them is extremely unreliable; and (4) in the region underlain by crystalline rocks it is probable that many drilled wells reported to have reached bedrock in reality are bottomed in large boulders. Reasonably full information was collected in the region around Kaukauna through the aid of the late J. J. Faust, a well driller. Many other drillers were interviewed, but only a few of them had any written records. Logs of all wells at Civilian Conservation Corps camps were secured, through the kindness of O. W. Potter, geologist for National Park Service, and some records of new wells were examined at the office of the State Board of Health. Figure 11 shows the bedrock surface in the vicinity of Black Creek, Outagamie County, where the drift is much thicker than it is in most of the district. The following logs of wells show the materials encountered at Black Creek and Gillett where the drift is also unusually thick.

Log of wells at Borden Company condensary, Black Creek, Wisconsin. Based on records and samples secured from C. H. Hauver and C. L. Green:

	Thickness (feet)	Depth (feet)
Red clay "hardpan" (red till?)	54	54
Sand	56	110
Clay, red, blue, black	58	168
Sand, gravel, and boulders	100	268
Sand	10	278
"Hardpan" (till)	53	331
Sand	2	333

	Thickness (feet)	Depth (feet)
Boulder (logged as "bedrock")	12	345
Sand and gravel	20	365
"Hardpan," blue and gray (till)	5	370
Gravel, coarse to fine, glacial; clay, red	76	446
Boulder	3	449
Sand, coarse, gray, cemented; layers of clay logged as "shale"	42	491
Clay, red, dolomitic, pebbly	7	498
Granite boulder	10	508
Gravel, pebbles of dolomite, sandstone, trap, porphyry	4	512
Granite, gneissic, pink	18	530
Dry hole in rock		

Log of wells at Gillett Canning Company, Gillett, Wisconsin. Based on records and samples from C. L. Green and J. J. Faust:

	Thickness (feet)	Depth (feet)
Till, red, dolomitic	70	70
Sand, pebbles of sandstone	20	90
Clay, red, dolomitic (lacustrine)	19	109
Sand, coarse to medium; gravel, fine	16	125
Clay, red, dolomitic (lacustrine)	60	185
Gravel, stony, coarse; layers of sand	40	225
"Hardpan" and undescribed drift	156	381
Sandstone, fine-grained, red, logged as "shale, white sandstone and limestone," apparently a cemented gravel or a breccia	31	412
Hornblende-mica-quartz schist	113	525
Small flow of water from schist; second well completed in gravel.		

Origin of rock basins.—Undoubtedly both the above deep wells prove the existence of basins in the surface of the bedrock. There are three possible origins for these: (1) preglacial river valleys which were tilted so that the headwaters became enclosed basins; (2) glacial erosion, and (3) plunge pools of glacial rivers. The small lateral extent of the basins seems to definitely exclude the first hypothesis. Exactly why glacial erosion should excavate these basins just below the border of the Lower Magnesium (Prairie du Chien) escarpment remains unexplained. Possibly the peculiar mixed rock and large boulders are not cemented gravel but are talus. An analogous rock basin 450 feet deep lies just east of Kaukauna at the foot of the Niagara escarpment. Glacial erosion is indicated by the relatively simple outline of the Platteville escarpment in Marinette County (Fig. 2). The outliers of Lower Magnesian in Mosquito Mounds east of New London show a slope toward the oncoming ice but no very extensive erosion. South of Shawano Lake the Lower Magnesian escarpment is irregular suggesting little glacial erosion of the bedrock. On many ledges more than one set of glacial striae occurs so that erosion by a later advance is demonstrably of small amount. Although many features of the unconsolidated drift survived overriding by a readvance of the ice, local disturbance was considerable. Transported masses of sand and gravel are common in the till. In many places disintegrated granite is found beneath the drift. The coarse-grained granites

have been much more altered than the finer-grained varieties some of which still preserve the glacial polish even where exposed to the atmosphere. Not a little of the breaking down of the granite seemingly occurred after glaciation, although such material would not be readily plucked. In conclusion, it seems that plucking was a necessary process for extensive glacial erosion of rock (Matthes, 1930). The origin of the rock basins of this area is still unsolved.

PREMORAINE GLACIATION

General considerations.—The drift outside the terminal moraine in Marathon and Langlade counties was assumed during the course of field work to be pre-Wisconsin. Subsequent study of the soil profile, however, has thrown doubt on such correlation. The profile is not that of a mature soil, although it reflects the results of poor subsoil drainage by yellow-brown and light gray mottling in the B_2 horizon. Poor subsoil drainage is the result of thin drift with high silt-clay content which lies on an impervious basement of crystalline rock. A few constructional features are present as was long ago recognized by Weidman (1907, p. 466-485, Pl. 2) who regarded such as his "Third Drift" and mapped a number of examples in Ts. 32 and 33 N., R. 9 E. in a region not surveyed in detail by the writer. The east-west border of this drift shown by Weidman suggests that the ice moved south. No Paleozoic chert was found. The recognizable erratics include abundant Keweenawan rocks, red felsite, red rhyolite, basalt, diabase, red sandstone, red shale, and some Huronian quartzite and iron formation. The Keweenawan, however, extends so far both to the northwest and the northeast of this area that exact determination of ice source is impossible. The writer was unable to check Weidman's drift border in northern Marathon County but drew a tentative line between fairly continuous till and the area of scattered erratics which trends northwest-southeast across Eau Claire River. The soils map of this area (Whitson, 1918a) shows the border of the Colby (Spencer) soils derived from drift farther south than this line (Pl. 10). South of this latter line the soils are mapped as Marathon or Gloucester, a series of soils largely formed from bedrock. Within the Marathon area, however, isolated areas of Colby were recognized. Weidman regarded most of the area south of his Third Drift as driftless. The writer agrees that it is in fact almost completely devoid of drift but does not agree that this means it was never glaciated. It is exceedingly difficult in a region of crystalline bedrock to distinguish glacially transported stones from residuals. Only the Keweenawan eruptives and sediments named above, the writer feels certain, are erratics. Using this criterion, as well as the evidence of the soils map, the writer

is confident that all of eastern Marathon County was glaciated. Probably the steeper slopes never had a very thick layer of drift, and this drift has been obscured by weathering or removed by erosion.

Arnott moraine.—The Arnott moraine of eastern Portage County is clearly the work of ice which came from the east as shown by its pebbles and boulders. It is older than the drift to the east as demonstrated by absence of deep kettles, the gullied sides of the ridge, and the deep soil profile. The soils map of Portage County (Whitson, *et al.* 1918) did not, however, recognize this weathering but included the soils as Coloma, the same as on the moraines farther east.

Correlation of extra-morainic drift.—Weidman (1907, p. 458-459) divided the area south of his Third Drift into Driftless and "Second Drift", the latter including the Arnott moraine. The writer was unable to check the triangular area of Second Drift in the vicinity of the Dells of the Eau Claire River, and it is not confirmed by the soils map. Evidence of westward-moving ice was observed in folding of gravel beds in the drift exposed in the brick pit at Ringle, an area called driftless by Weidman. Probably the ice which deposited the drift outside the Wisconsin terminal moraines came both from the north and the east. Whether these two ice invasions were simultaneous or of different ages is uncertain, and decision awaits further study of soil profiles. Evidence now available does not suggest a wide difference in age because both display some constructional features. Leverett's (1932, p. 22-23) maps published by Antevs (p. 544, 646) correlated all this drift as Illinoian. This correlation is uncertain because the area is so far removed from the type locality of that drift. The moderate amount of erosion and weathering at most localities distinctly favor an Iowan age rather than Illinoian or older. The possibility remains, however, that there is more than one age of drift in Marathon County and that the patchy drift of Weidman's "driftless" district is older than either the northern drift of Langlade County or the Arnott moraine. The vague and irregular border of the thin drift does not suggest a boundary due to glacial deposition.

INTERGLACIAL INTERVAL

Owing to the low relief of most of northeastern Wisconsin and the scarcity of deep excavations, no contact has ever been observed between the Wisconsin drift and the extramorainic till. It is evident, however, that the time between the two was at least long enough for all residual ice to melt and that there is a distinct difference in amount of weathering under comparable topographic conditions.

WISCONSIN GLACIATION

General statement.—The recognized Wisconsin drift of northeastern Wisconsin is readily divisible into two distinct substages. The divisions of the Wisconsin stage have been given various names and numbers. The Iowan is included in the Wisconsin by Kay and Leighton but is placed

TABLE 1.—*Substages of Wisconsin glaciation*

	Leverett, 1929	Kay and Leighton, 1933	Thwaites, 1934	Thwaites, 1941
Late	4th, 5th	Mankato	4th, 5th	Mankato, Valders
Middle	3d	Cary	3d	Cary
Early	1st, 2d	Tazewell	2d	Tazewell
		Iowan	1st	Iowan

in the Illinoian by Leverett. Table 1 explains the several proposed classifications the respective merits of which will not be discussed here. Tripartite division into Early, Middle, and Late Wisconsin is not broad enough. On the other hand, the numerical system is easy to remember but is too inflexible in case errors in sequence are discovered. Geographic terms with no fixed order are, therefore, preferable. The type locality, Mankato, Minnesota, chosen by Kay and Leighton for the post-Cary substage is separated from the "red till" of Alden (1918, p. 310-324) by so vast an expanse of unsettled, little-known territory that equivalence is far from assured. Pending further study of northwestern Wisconsin and northern Michigan it seems advisable to adopt a new designation for the red till in northeastern Wisconsin, and the name "Valders" here proposed is based on the exposure of the two tills with associated crossing striae in the quarry at Valders, Manitowoc County, Wisconsin (Alden, 1932, p. 44). No definite time relation to the Mankato of Minnesota and Iowa is then implied. Correlation of the gray till of northeastern Wisconsin as Cary is much better assured as field tracing of the moraines (Fig. 1) is complete. The Cary appears to be the type which was originally proposed for the Wisconsin drift (Thwaites, 1927, p. 64).

Outer moraine.—The Outer moraine of the Green Bay lobe (Fig. 1) is a continuous ridge with the exception of a few narrow drainage outlets. It has been traced to a connection with Alden's Johnstown moraine near the south line of Waushara County. The relation of moraines at that point suggests, however, that the Outer moraine has been overridden by the Johnstown moraine. If so, the Outer moraine may correlate with the Brooklyn moraine of southern Wisconsin (Alden, 1918, p. 184-186,

FIGURE 12.—*Conditions at time of formation of Outer (Brooklyn?) and Parrish moraines*

261-262). Alden tentatively ascribed the Brooklyn moraine to the early Wisconsin or what is now termed Tazewell. Investigations by the present writer show that such correlation is incorrect and the Brooklyn is an older Cary moraine. Evidence consists of (1) the occurrence of pitted outwash in the space between the Brooklyn and Johnstown moraines (Thwaites, 1926, p. 317); (2) the lobation of the Brooklyn moraine is nearly the same as that of the Johnstown, whereas on the Wisconsin-Illinois border there is a marked discordance between Tazewell and Cary ice directions; and (3) a dual maximum of the Cary was disclosed by drilling in the vicinity of Prairie du Sac, Sauk County, where the Johnstown moraine crosses Wisconsin River (Thwaites, 1926, p. 317).

Parrish moraine.—The Parrish moraine of the Langlade lobe is contemporaneous with the Outer moraine (Fig. 12). The two moraines merge in eastern Langlade County into an indivisible complex; outwash from both moraines mingled in the Antigo Flats as shown by slopes and pebble counts; drainage channels from the Flats through the Green Bay moraine to the low ground along Wolf River do not exist (Pl. 1); much pitted outwash lies within the area of Green Bay drift which by its low content of dolomite pebbles (Fig. 4) shows it was derived from the Langlade lobe during essentially contemporaneous wastage of the two lobes; no nonpitted outwash occurs along the contact of the two lobes as would be expected if one had retired long before the other; and some of the Green Bay outwash in the Antigo Flats was terraced by drainage from the Langlade Lobe. Distribution of striae and drumlins (Pl. 10) indicates that the Langlade lobe at its maximum extended much farther southeast than does its drift at the surface today.

Drainage at Maximum of Cary Substage: Outwash was deposited by streams which left the Outer and Parrish moraines forming the Antigo Flats in the re-entrant angle between the two lobes. Waters from this area escaped to Wisconsin River via Eau Claire River forming the Dells where a divide was crossed. Weidman (1907, p. 498-501) held another view of the Antigo Flats, stating:

> "This tract is generally characterized by gravelly loam, with some boulders, with an occasional depression or sag, and by swells of coarse drift. This plain about Antigo has the characteristic features of an old alluvial plain over-ridden by the Third Drift—. It is difficult to separate the outwash from the Wisconsin ice, but material of the latter sort is believed to be insignificant in amount. The slope of this gravelly plain in front (north) of the Green Bay moraine in this vicinity is downward to the south, thus sloping downward towards the moraine, a condition of slope inconsistent with the theory that any important part of it was deposited by streams issuing from the Green Bay lobe."

Evidently Weidman was misled by several phenomena: (1) slight depressions in channels of abandoned streams simulate kettles; (2) there

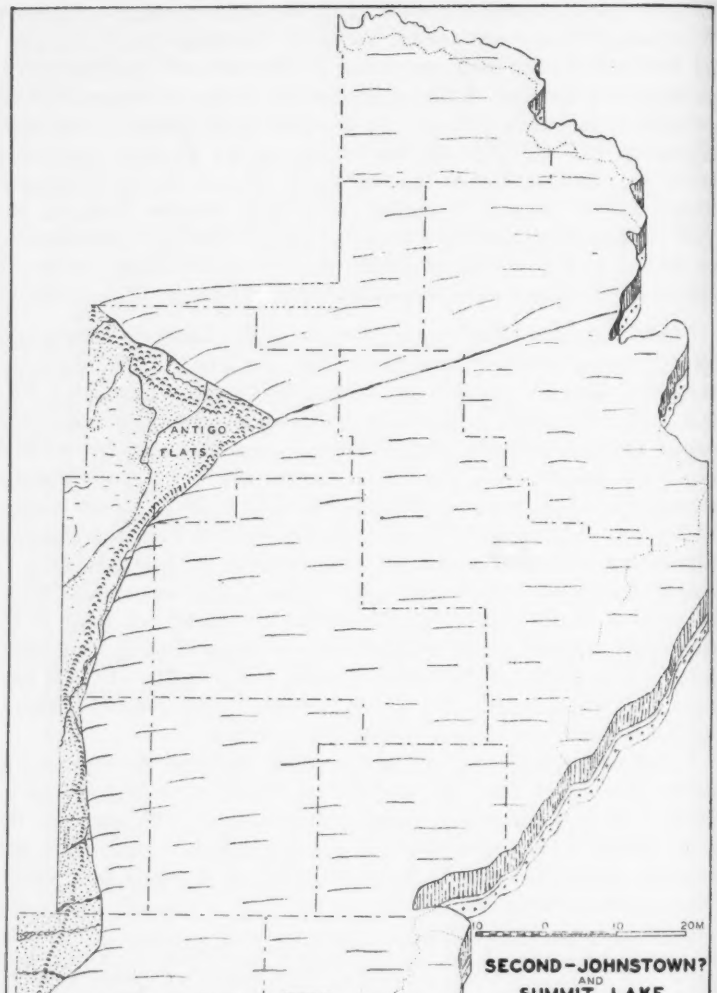


FIGURE 13.—*Conditions at time of formation of Second (Johnstown?) and Summit Lake moraines*

are islands of till which are almost buried; (3) some ice-rafted boulders occur; (4) the weathered zone or soil profile which explains the greater amount of silt and clay near the surface is deceptively like till; (5) apparently Weidman did not visit the eastern part of the Flats (Pl. 2, fig. 1; Pl. 10) where there is a marked slope north from the Green Bay moraine; and (6) Weidman could not have examined deep cuts which show dolomite pebbles from the Green Bay drift (Fig. 4). Possibly the small amount of outwash along the Outer moraine for a considerable distance east of Wausau led Weidman to deny the abundance of glacio-fluvial deposits. Throughout this area meltwaters must have traveled subglacially to the Plover River outlet farther south (Fig. 13; Pl. 10). Weidman (1903, p. 18, 62) reached this unusual opinion in his first report on soils.

Second moraine.—The Second moraine of the Green Bay lobe appears to be correlative with the Summit Lake moraine of the Langlade lobe. Opening of a drainage line along the course of Plover River (Fig. 13) allowed the meltwaters of the Green Bay lobe to escape southward to the Plover River gap whereas waters from the Langlade lobe still crossed the Antigo Flats. These northern waters caused terracing. South of Plover River were several avenues of escape through gaps in the Outer moraine. Outwash between the two moraines was deposited over much stagnant ice particularly in long strips just inside the breaks in the Outer moraine. Possibly these were subglacial channels during the formation of that moraine, and thus the ice lasted longest in the depressions. The large outwash plain in Waushara County is known as Great Prairie. The marked bend in the course of the Second moraine in Waushara County is probably explicable by a southeastward extension of the Arnott moraine. The highest elevations along this part of the Second moraine occur at this angle. The regular outline of the Second moraine elsewhere points to a major readvance of the ice. Possibly the Second moraine correlates with the Johnstown moraine, but if the Outer moraine is Johnstown then this is the Milton moraine.

Elderon morainic system.—The complex system of three to five moraines inside the Second moraine is here termed the Elderon morainic system. It may embrace the Milton moraine of the region south of this area. If so, the Lake Mills system is also included. Recessions, rejuvenations, and readvances occurred in rapid succession, for all the associated outwash is highly pitted. Figure 14 shows that the corresponding Langlade ice border was at the Elcho moraine. Drainage from that moraine was still south across Antigo Flats, but the Green Bay ice drained along its margin south to the gap at the State highway west

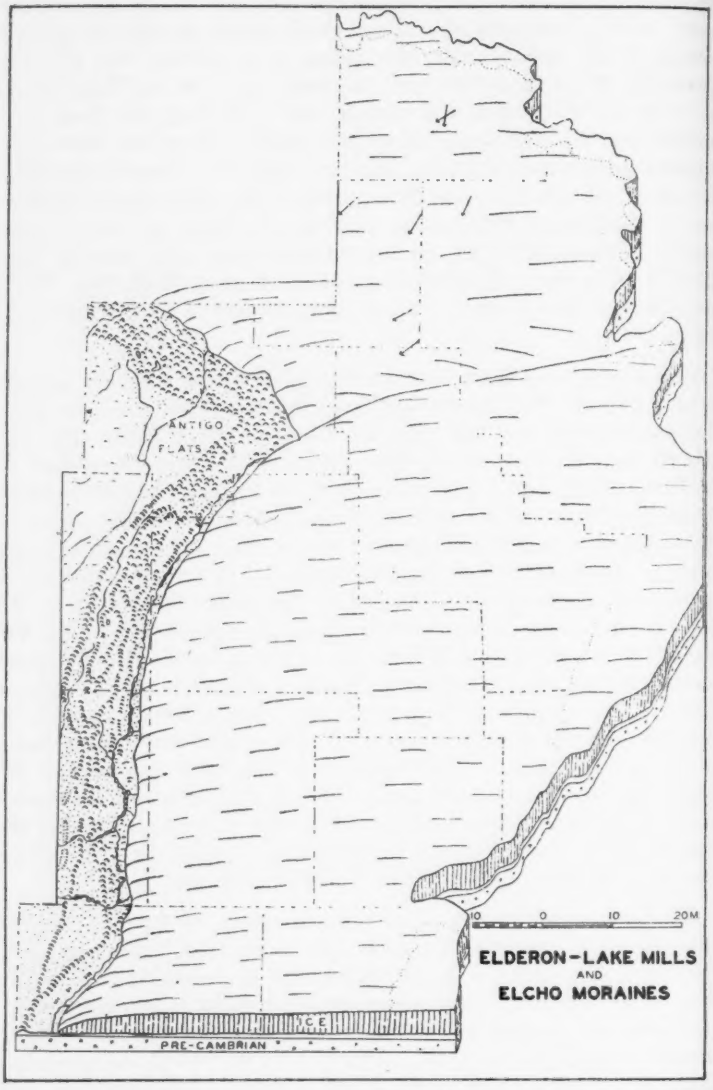


FIGURE 14.—*Conditions at time of formation of easternmost Elderon (Lake Mills) and Elcho moraines*

of Waupaca. Another outlet farther south in the same county seems to have been used at this time as evidenced by erosion channels across Great Prairie. In Waushara County drainage was south to Lake Wisconsin which then extended east of the Johnstown moraine because the

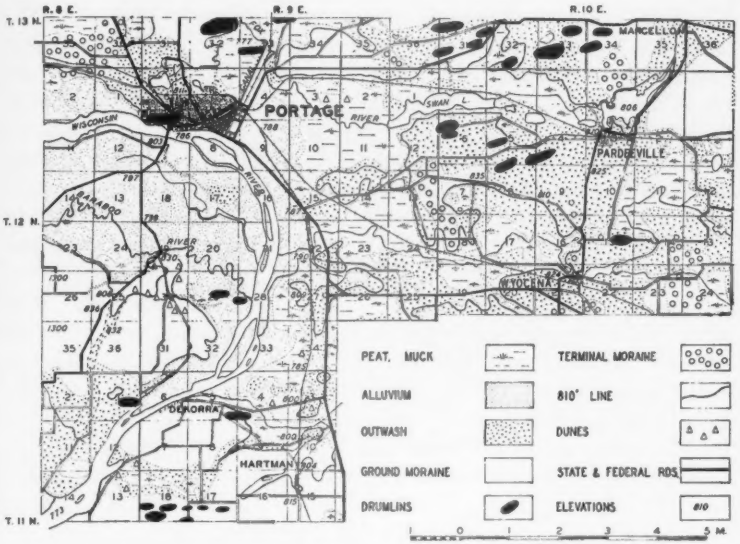


FIGURE 15.—*Outlet of Lake Oshkosh near Portage, Wisconsin*

Based on field work by W. C. Alden, K. S. Kurtenacker, J. M. Wanenmacher, and the writer, soils map of Columbia County, U. S. War Department surveys, railroad surveys, U. S. Geological Survey maps, and aerial photographs.

ice front had not yet cleared the east end of the Baraboo Bluffs (Fig. 15). Much of the outwash deposited along the Green Bay ice margin in Langlade County at this time is unusually coarse and difficult to distinguish from kame gravels of the moraines.

High outwash terraces.—Following the abandonment of the eastern-most Elderon moraine the ice margin melted back rather rapidly, and large amounts of intensely pitted outwash accumulated along it. The eastern margin of the "High Plains" (Pl. 9, fig. 2) is very irregular and is devoid of a till moraine. It is shown on Plate 10 by a line and cannot be traced with assurance very far north into Shawano County. East of this line several hills of sand and gravel rise almost as high as the level of the plains which were still graded to Lake Wisconsin.

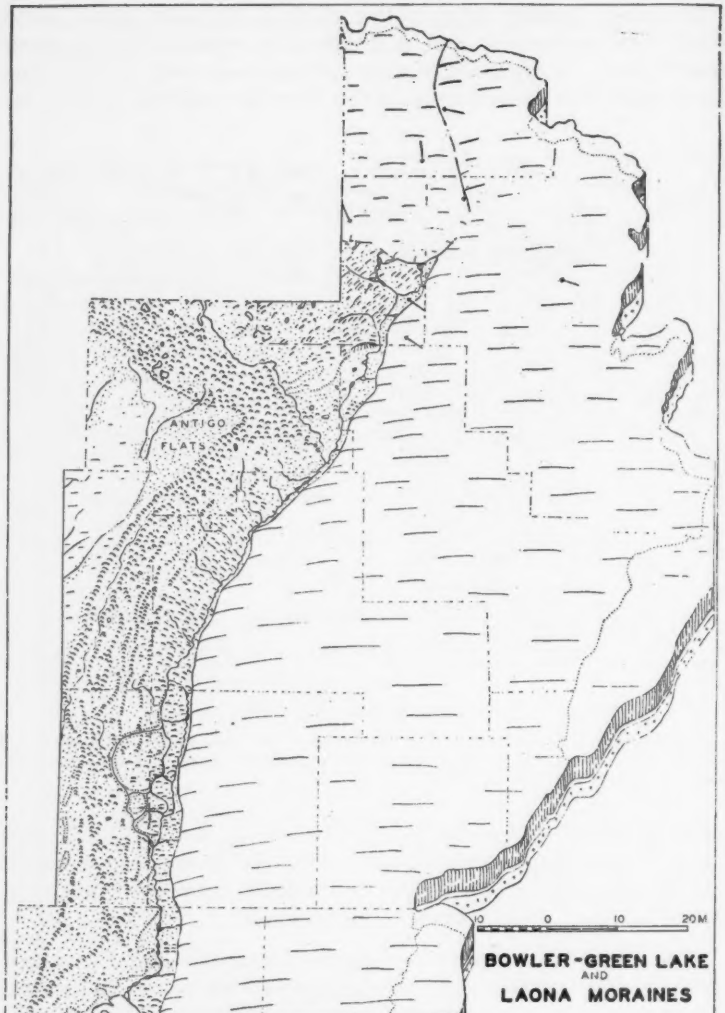


FIGURE 16.—*Conditions at time of formation of Bowler (Green Lake) and Laona moraines*

Striae ascribed to this readvance indicated. Meltwaters escaping south along ice front to beginnings of Early Lake Oshkosh in Waushara County eroded complex system of drainage channels in High Plains.

It is suggested that at this time melting so reduced the slope of the margin of the glacier that forward motion ceased (Thwaites, 1934, p. 18). The isolated hills would then be classed as moulin fillings within the stagnant ice.

BOWLER MORAINIC SYSTEM

Figure 16 shows the ice margin for the time of deposition of the Bowler and Laona moraines. The Bowler system consists of two to four separate ridges. It seems to join the Green Lake moraine of southeastern Wisconsin. The relation of the Bowler moraines to the drift of the Langlade lobe plus the evidence of outwash buried below till and of crossing striae clearly demonstrate a marked readvance of the ice after a period of recession to an unknown distance but for a time insufficient to melt all residual ice remnants. This readvance seems to have made the Green Bay ice invade much of the former Langlade territory leaving a shrunken Langlade lobe with a slightly different direction of movement from that which prevailed before. If any interlobate moraine once existed, it was thus destroyed. Drainage was to the south along the ice front but through such a complex of drumlins and earlier moraine that exact tracing is difficult. The marked erosion valley west of Lakewood (Pl. 10) was formed at this time. The terracing due to many changes in outlet and melting of obstructing ice was complex. Many earlier moraines were either buried or eroded depending on the local grade of the streams.

FORMATION OF LOW PLAINS

At this time the ice margin cleared the east end of the Baraboo Bluffs (Fig. 15), and the level of waters impounded in front of the ice fell from about 980 to about 840 feet. The initial outlet was the channel along the State highway at the foot of the bluffs whose bottom now lies at elevation 832. Outwash graded to this new level formed the Low Plains about 150 feet below the high level outwash to the west. The lower part of the escarpment between the two outwash levels carries some local varved clays for instance at Richford, Waushara County. During the formation of the low terraces the high-level outwash plains were dissected by two distinct types of valleys. The spectacular terraces near Amherst, Portage County, are outlined on Plate 10. The valley now occupied by Waupaca River continues northeast to Northland through a low col not now followed by any stream. East of Amherst are several channels which branch and reunite in a complex manner. Some of these hang above the levels of those they join. Evidently they record shifting drainage along the ice front, and all of them were not occupied at one time. A second class of valleys indent the face of the escarpment. These

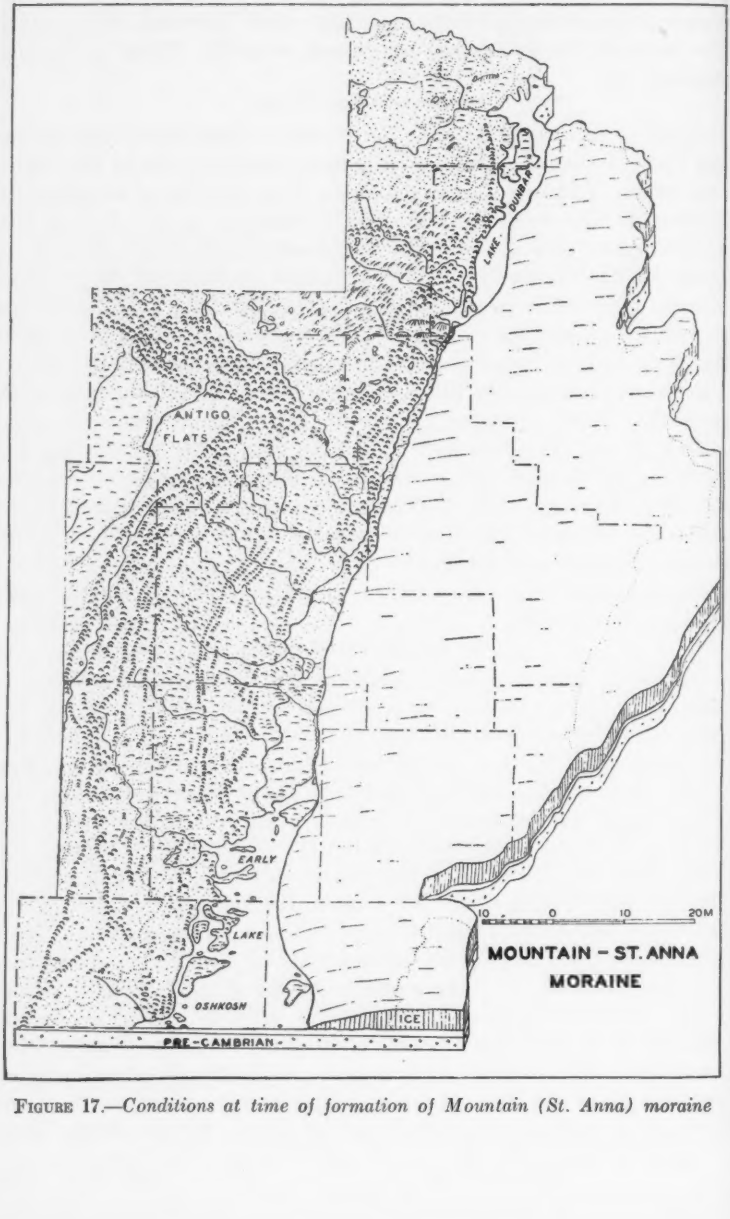


FIGURE 17.—*Conditions at time of formation of Mountain (St. Anna) moraine*

are now occupied by small, spring-fed streams which look much too weak to have formed such large valleys. There are no alluvial fans on the low plains opposite the mouths of valleys; if formed by post-glacial erosion, it seems as if there should be such fans. Drainage from ice

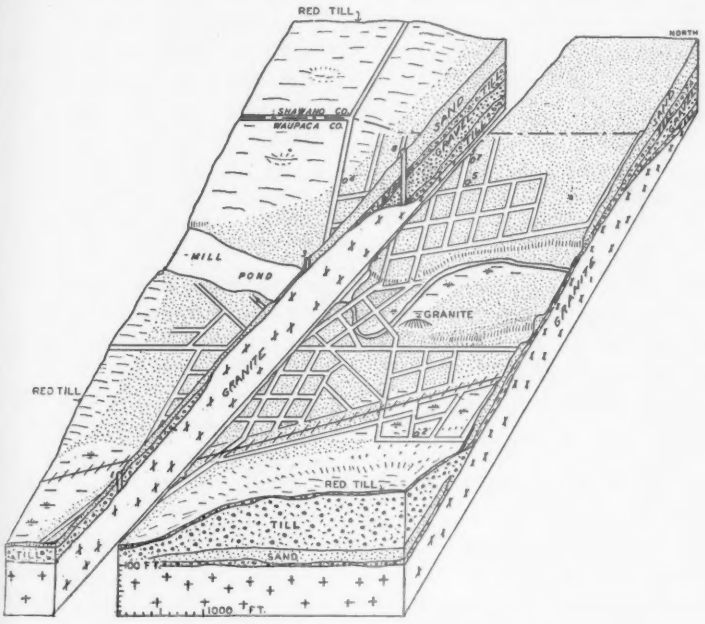


FIGURE 18.—*Vicinity of Marion, Wisconsin*

Block diagram based on test borings for village water supply by Layne-Northwest Company.

blocks buried in the high-level outwash plus normal precipitation possibly caused rapid erosion of these valleys while streams were still flowing on the Low Plains to carry away the material thus removed.

MOUNTAIN MORAINIC SYSTEM

After recession of the ice margin from the easternmost Bowler moraine, at least three distinct ice border positions were occupied prior to the condition represented in Figure 17. These earlier moraines of the Mountain system appear to correspond to the Waupun and Rush Lake moraines which were mapped farther south by Alden. The distance between moraines increases somewhat toward the south. South of central Shawano County all these moraines were covered by the till of the Valders sub-

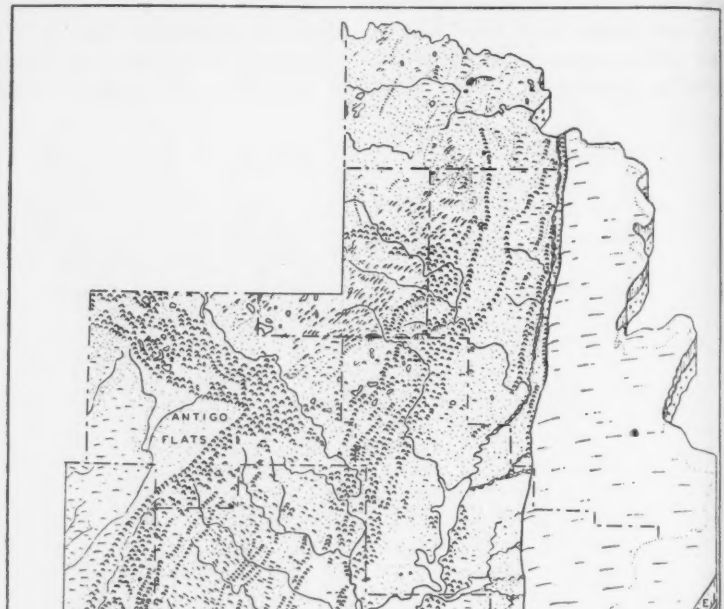


FIGURE 19.—*Conditions at time of formation of Athelstane moraine and maximum of Early Lake Oshkosh*

stage and much subdued. In places they were entirely obliterated by later lake deposits. Marginal drainage produced the channels through the Low Plains of Waushara County and truncated the ends of drumlins at Poysippi. Some low-level outwash terraces were formed, but drainage was into Lake Oshkosh, although probably at a slightly lower level after the outlet at Dekorra (Fig. 15) was freed of ice. The water level at the outlet then did not reach 822 feet as there was no overflow along the railroad south of Hartman. The ice front shown in Figure 17 appears to correlate with the St. Anna moraine to the south. That each moraine is the product of a rejuvenation is shown by buried outwash. In fact, it is probable that the irregularity of outline of some of the areas mapped as moraine indicates that some areas of overridden pitted outwash were erroneously included.

MARGINAL LAKES

In Marinette and Florence counties lake clays have been found under outwash sand and gravel all along the outside of the easternmost Mountain moraine from the gap in McCaslin Mountain north to the State line. On the county road south of Dunbar (Sec. 36, T. 36 N., R. 18 E.) this clay underlies the till of the moraine and rests upon an older outwash. Apparently the readvance blocked flow past the east end of the mountain and formed a body of water for which the name Lake Dunbar is proposed. This long, narrow lake was filled with outwash so that little appears on Plate 10. Early Lake Oshkosh was recorded in many localities by varved clays and beach deposits, many of the latter somewhat higher than later post-Valders beaches. Test borings at Marion, Waupaca County, shown in Figure 18 demonstrate that another marginal lake was formed by the Mountain readvance, surface evidence of which was destroyed by later burial under outwash.

ATHELSTANE MORAINIC SYSTEM

No figure is presented to show conditions when the ice stood at the westernmost moraine of the Athelstane system. At this time the extensive outwash plain north of Shawano Lake was deposited with its south end in Early Lake Oshkosh at an elevation close to 850 feet. The outer Athelstane moraine is buried under Valders till from a point west of Crivitz south. In Shawano County it is still a very striking feature, however (Pl. 4, fig. 1). Farther north destruction has been more complete. To the south it appears to curve eastward and pass well to the north of Lake Winnebago. At this time the Duck Creek eskers were formed (Fig. 10). The deltas near Green Bay suggest that the lake level fell to around elevation 800 feet by this time. The outwash buried

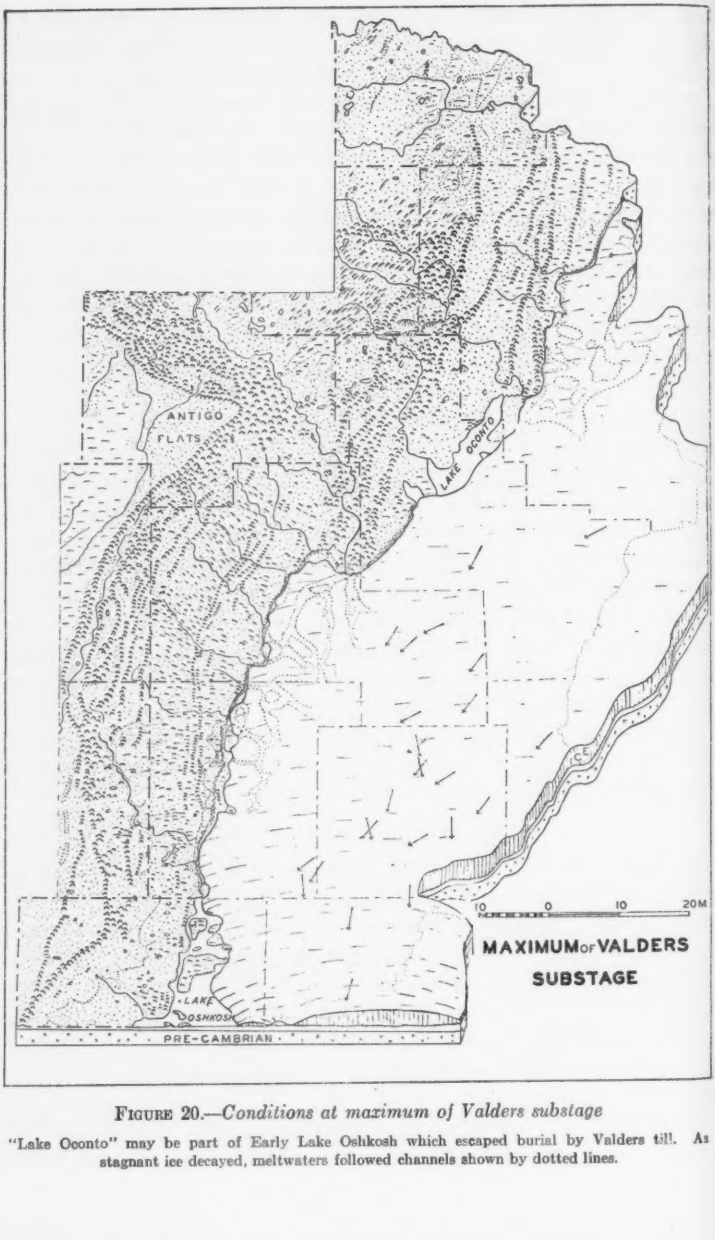


FIGURE 20.—*Conditions at maximum of Valders substage*

"Lake Oconto" may be part of Early Lake Oshkosh which escaped burial by Valders till. As stagnant ice decayed, meltwaters followed channels shown by dotted lines.

under red till east of Pulaski seems to indicate a level which was possibly even lower.

TILTING OF EARLY LAKE OSHKOSH BEACHES

The maximum extent, although not the maximum elevation, of Early Lake Oshkosh is indicated in Figure 19. Trainer (p. 37-43) reported a beach at 900 feet just south of this area near Berlin. This conclusion does not check with the evidence of outwash plains near Portage (Fig. 15). Extensive exploration in the intervening area carried out by K. S. Kurtenacker, P. J. O'Neil, and R. F. Sanderson under the direction of the writer failed to confirm any beaches above 850 feet elevation.[1] Alden (unpublished field map) apparently regarded the high sand deposits at Berlin as dunes, a view in which the writer concurs, but some may have been deposited in local ice margin pools. On the soils map most are classed as Coloma (Whitson, 1929b). Moreover, there is no divide southwest of Berlin high enough to make a dam for the supposed maximum stand of the lake. High isolated hills within the area supposed to have been submerged show no trace of beach deposits. However, the varved clays at Richford, Waushara County, are at an elevation of about 915 feet, and in many places beachlike gravels occur at elevations well above 850 feet. Almost any system of tilted beaches may be made from these facts. However, no continuous beach can be traced, and known occurrences of lake sediments are readily explained by local pools along a retreating ice front. Possibly the deltas and varved clays up to 860 feet north of Suring, designated on Figure 20 as Lake Oconto, are the northeasternmost deposits of Early Lake Oshkosh. Were more known of the varve succession, this correlation could probably be either proved or disproved. If they are deposits of the earlier lake formed during the Cary dissipation, then tilting would have to be recognized and would amount to 60 feet in almost 140 miles, or less than 6 inches to the mile. The movement might, however, have been confined to the northern part of the basin.

DISAPPEARANCE OF CARY ICE

After the ice margin left the easternmost Athelstane moraine, there is no record of another Cary readvance in Wisconsin. Very likely the ice became stagnant, although definite proof of this is lacking. A glacial lake in the Green Bay lowland is demonstrated by varved clay a short distance west of Suamico which was disturbed and buried by the Valders ice. This lake was an extension of what the writer terms Early Lake Chicago (Thwaites, 1934, p. 87, 98-99) but which Martin (1932, p. 454) named Lake Milwaukee. The clays appear to be the same as those

[1] Reports on file in Library of University of Wisconsin, Science Hall, Madison, Wisconsin; also a master's thesis by R. W. Ellis in main library.

below the Forest Bed of Manitowoc County, but this correlation was not checked by a varve curve. It seems certain from the position of the organic remains at the Forest Bed near Two Creeks, Manitowoc County (Wilson, 1932; 1936; Alden, 1932, p. 43; Thwaites, 1934, p. 83-84) that the Cary ice margin receded so far that the Straits of Mackinac were opened. The only organic remains at this horizon in the area here described lie in the district between Appleton and Oshkosh. Logs and peat are reported in many wells in this vicinity. Lawson (1902) described exposures of similar vegetal material which no longer exist. The writer collected a spruce log flattened by ice pressure from varved clays on the east shore of Fox River in SE¼SE¼ sec. 3, T. 20 N., R. 17 E. In his second paper Wilson (1936) states: ". . . though the fossil plants have a decided boreal character they also represent the pioneer organisms of denuded areas uuder certain climatic conditions and are therefore not reliable indicators of a severe climate." The soil profile developed at Two Creeks is infantile and demonstrates the brevity of the interval during which the land was free of ice. None of the numerous exposures of the contact of the Valders till with older glacial drift in the area under discussion shows any trace whatever of weathering or organic growth.

VALDERS GLACIATION

General considerations.—The readvance of the ice sheet shown in Figure 20 is here named Valders. Drift of this age is readily distinguished from the Cary till because it is red and contains much more clay. This lithologic criterion is not easily applicable in the northern part of the district. The red color is best shown where the Valders till overlies older lake clays as it does in the Fox and Wolf valleys. Farther north the till rests on outwash sands and older gray till and is light pink in horizon C of the soil profile. The soil profile varies considerably throughout the district. Where the subsoil is relatively porous the A horizon, a gray silt, is deepest, locally over 2 feet; this explains why the soil was erroneously called Miami on some of the older soils maps. In the B horizon the color is a deep brownish red, a poorly developed ortstein, and is readily confused with the same horizon in soils on the Cary gray till. Where the subsoil drainage is poor, both the A and B horizons are thin, the former in few places over a foot thick (Kellogg, p. 75-83). On the later soils maps this shallow profile is what is called Superior, whereas the more rolling and better drained soils on red till are termed Kewaunee.

Origin of red color.—Chamberlin (1877, p. 219-228) regarded the red till of northeastern Wisconsin as a lake deposit. Alden (1918, p. 310-324)

disproved this view and showed it to be a glacial deposit. The writer confirmed Alden's interpretation by (1) evidence of disturbance of underlying deposits, and (2) the termination of the red till in many places on ground higher than that outside its area. Both mechanical and chemical analyses (Figs. 5, 7) show that it is in fact very similar to lake clays of the same region, but field inspection shows many more boulders and pebbles. Alden's explanation of the source of the till by plowing up older lake clays which were colored red by glacial drainage from the Lake Superior region appears adequate, for the Valders till northwest of the lake clay region is much less red. As the Valders ice came from a source decidedly west of that of the Cary invasion, some red material may have been contributed directly by the ice, but this appears to have been of minor importance (Thwaites, 1934, p. 84).

Lobation of the Valders ice.—Figure 20 shows what is known of the maximum extent of the Valders ice. Owing to the thin drift indicating a very short period of ice occupancy, the Valders ice left few definite traces of movement. Some striae which are out of harmony with the Cary motion as recorded in drumlins are definitely ascribed to the Valders invasion. The same may be said of the drumlins south of Shawano and the new tails built onto the lee sides of some other drumlins (Fig. 6).

Drainage of Valders ice.—The absence of an end moraine at the margin of the Valders red till indicates that the ice did not long remain at this position. The tracing of the edge of the Valders glaciation is uncertain north of Shawano County because the till is not readily distinguished. The writer traced this boundary with reasonable satisfaction up to the north line of Oconto County, but across Marinette County it was very difficult to follow. The line given on Plate 10 may be much too far southeast. It certainly does not agree with the line drawn by Leverett (1929, Pl. 1). The absence of any recognizable moraines within the Valders area, combined with the meandering courses of many eskers (Pl. 5, fig. 1), indicates that the ice became stagnant immediately upon reaching its maximum extent. Only a little non-pitted outwash occurs along the margin although it appears from the relation of red till to kettles that all the buried ice masses had disappeared before the coming of the Valders ice. In the southern part of the area a consistent marginal drainage slope could be made out. North of the westward salient near Gresham, where the border of the Valders drift reaches an elevation close to 1000 feet, mapping of marginal drainage is difficult. With the ice border so high in Shawano County why are no unequivocal records found of lakes impounded by the Valders ice? The area

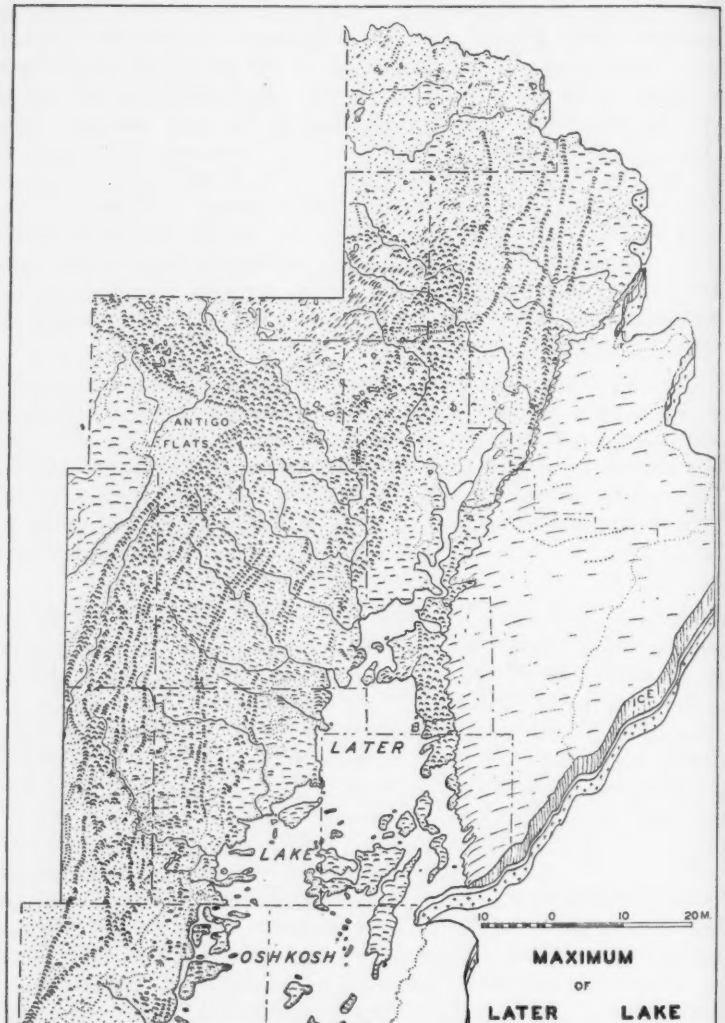


FIGURE 21.—*Conditions at maximum of Later Lake Oshkosh*

Position of frayed ice margin can only be conjectured from distribution of beaches of Later Lake Oshkosh.

mapped as Lake Oconto has an apparent elevation of only 860 feet thus raising the question of its true age. If there were any marginal lakes along this part of the border, it is possible that their record has been obscured by later outwash and dune formation. Farther north almost no definite evidence of marginal drainage could be discovered despite a thorough examination of both the surface and the aerial photographs. Was the border so inconstant in position that it left no trace, or is it much farther west than is shown on Plate 10? In view of the known brevity of the Valders maximum it seems clear that very soon the waters began to flow through the plexus of channels shown on Figure 20 in dotted lines and as Valders outwash on Plate 10.

Later Lake Oshkosh.—Figure 21 shows the maximum determinable extent of Lake Oshkosh which existed during the recession of the Valders ice. The ice front is shown as very ragged because it was formed by melting back of stagnant ice and not by a rejuvenation and readvance. As pointed out previously, the beaches of Later Lake Oshkosh cannot be traced continuously. The amount of sediment was considerable in some places, and the clays shown in Fig. 2 of Plate 8 accumulated, but over much of the lake bed deposition was slight. The most easterly known beach deposit is on top of the hill at the gravel pit in sec. 5, T. 21 N., R. 18 E. north of Little Chute. The elevation there was estimated at 815 feet. Still later the pass through the Niagara escarpment at Sturgeon Bay, Wisconsin, east of the area here described was freed of ice, and Later Lake Oshkosh disappeared. Outwash was then deposited as low as elevation 630 feet.

Tilting of Later Lake Oshkosh: Only a few accurate determinations of the maximum level of Later Lake Oshkosh could be made. In sec. 31, T. 24, N., R. 15 E., near Bear Creek a railroad survey places the top of a bar at 829 feet. This is 82 miles from the outlet at Dekorra. Beach gravels at the northeast corner of sec. 36, T. 21 N., R. 16 E., west of Appleton, are at about the same elevation as determined by the United States Geological Survey. This point is 73 miles from Dekorra. The water level in the outlet is difficult to fix. During the interval between the Cary and Valders invasions the course of Wisconsin River at the point may have been raised by deposition of sand. Furthermore, when a large portion of the ice front drained through this one outlet, the water level was higher than when the Pardeeville outwash plain (Fig. 15) was deposited by waters from a relatively short section. These factors combined might readily have kept the water in the narrow outlet as high as 820 feet. If this is the case, the tilt indicated would be so slight that it is below the limit of error in determining water planes. If, how-

FIGURE 22.—*Conditions at time of Lake Algonquin*

ever, the water stood at an elevation of only 800 feet the tilting would be about 4 inches to the mile. It would seem, therefore, that there is little definite proof of tilting in the region considered and that postglacial earth movements were probably confined to the region farther to the north.

WIND DIRECTIONS: The positions and forms of the beaches and bars of Later Lake Oshkosh give a clue to the direction of winds during the wasting of the Valders ice. With the presence of an ice sheet on which the atmospheric pressure would almost always be relatively high, we should expect air drainage and, therefore, winds considerably different from those of today. Among the better-developed beaches amid the islands of the southern part of the lake basin the majority of deposits require a north, northeast, or east wind for their formation.

Terrace formation.—As already pointed out it is difficult to draw any sharp line between glacial outwash terraces and postglacial terraces. The Clintonville delta was certainly formed in a Lake Oshkosh at an elevation close to 825 feet and with icebergs still present although the source of water in Pigeon River is not evident. One cause of a large flow in the streams of late glacial time was the slow melting of buried ice masses. These would tend to give a much more equable flow to the streams than they have had either before or since. However, much terracing occurred after all ice blocks had melted, for there are no kettles in many terraces. For a long time after glaciation much of the present swamp area must have been open water, but subsequent accumulation of organic debris has obscured the evidence of old shore lines.

Sand dune formation.—Formation of dunes took place mainly in the lake beds and evidently lasted a long time. A large portion was made by westerly or southerly winds (Pl. 9, fig. 1) like those of today. Most of the dunes were covered with vegetation before white men entered the region; many have been revived by cultivation and forest fires.

LAKES ALGONQUIN AND NIPISSING

After the Valders ice had left the region, the waters of Lake Michigan basin stood at the level known as Lake Algonquin (Fig. 22). In view of the poor evidence of this lake and its successor, Lake Nipissing in this area, it is not necessary to elaborate on the history which has been worked out elsewhere (Leverett and Taylor, 1915). Clearly earth movement has taken place since the formation of the Algonquin beach giving it a slope of about half a foot to the mile rising toward the northeast. A. L. Hillis, city engineer at Marinette, reports many oak logs buried in the lake or stream sands below the Algonquin beach. A sample was determined as white oak by Professor R. H. Denniston.

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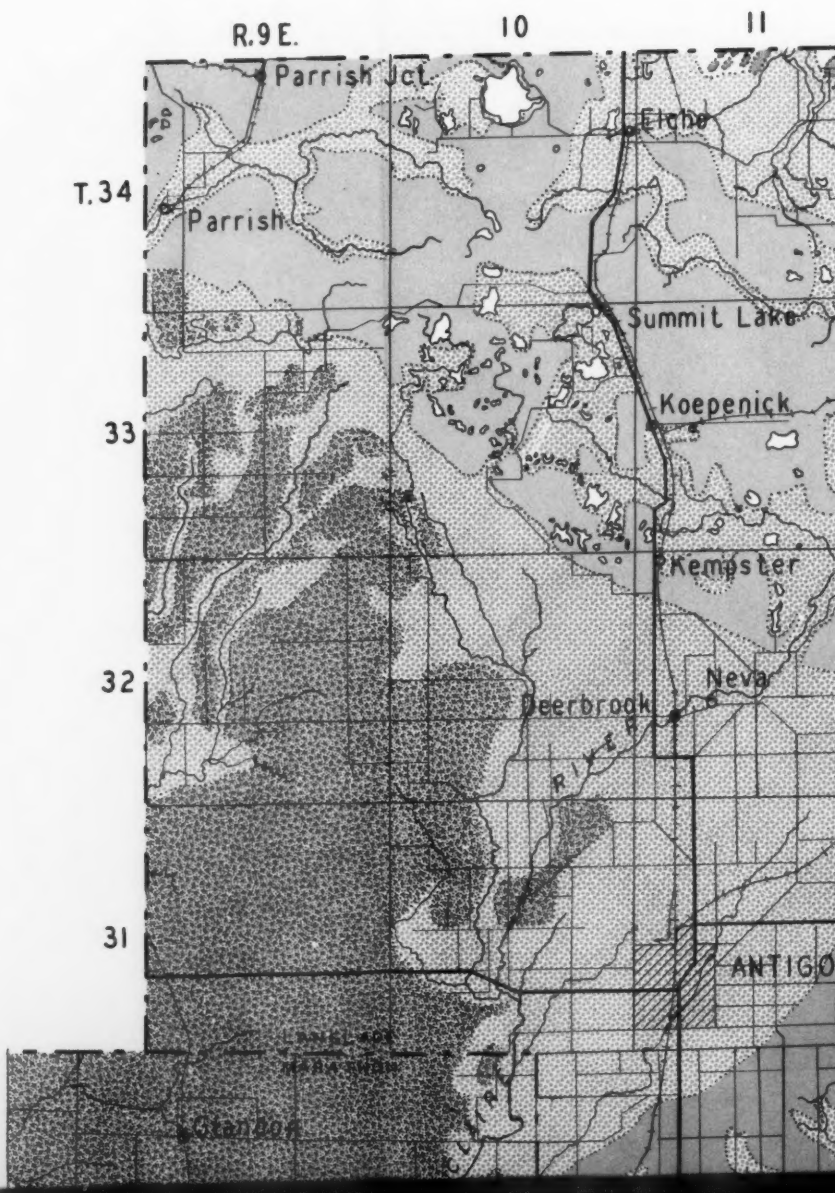
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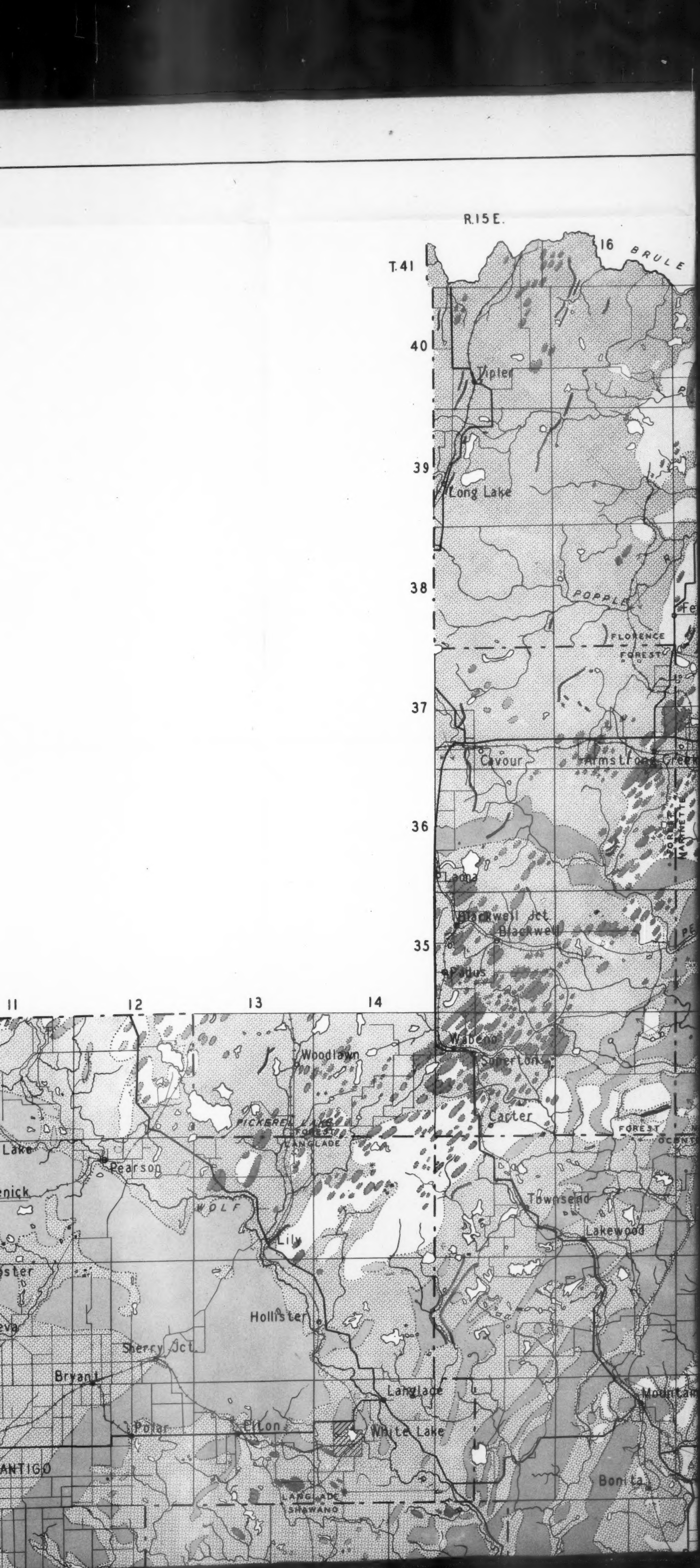
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Blackwell Jct
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Wabeno
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Copperton
Carter
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MARINETTE
OCONTO
CALDRON FALLS RESERVOIR
Townsend
Lakewood
HIGH FALLS
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Langlade
Mountain
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LAKE
Bonita
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FLORENCE

MARINETTE

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Pembine

Beecher

RIVER

Amberg

Cedarville

Athelstane

Wausaukee

R22 E.

RIVER

Packard

McAllister

Wagner

Middle Inlet

Goll

LAKE NOQUEBAY

Miles

PESHTIGO

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Porterfield

Bagley Jct

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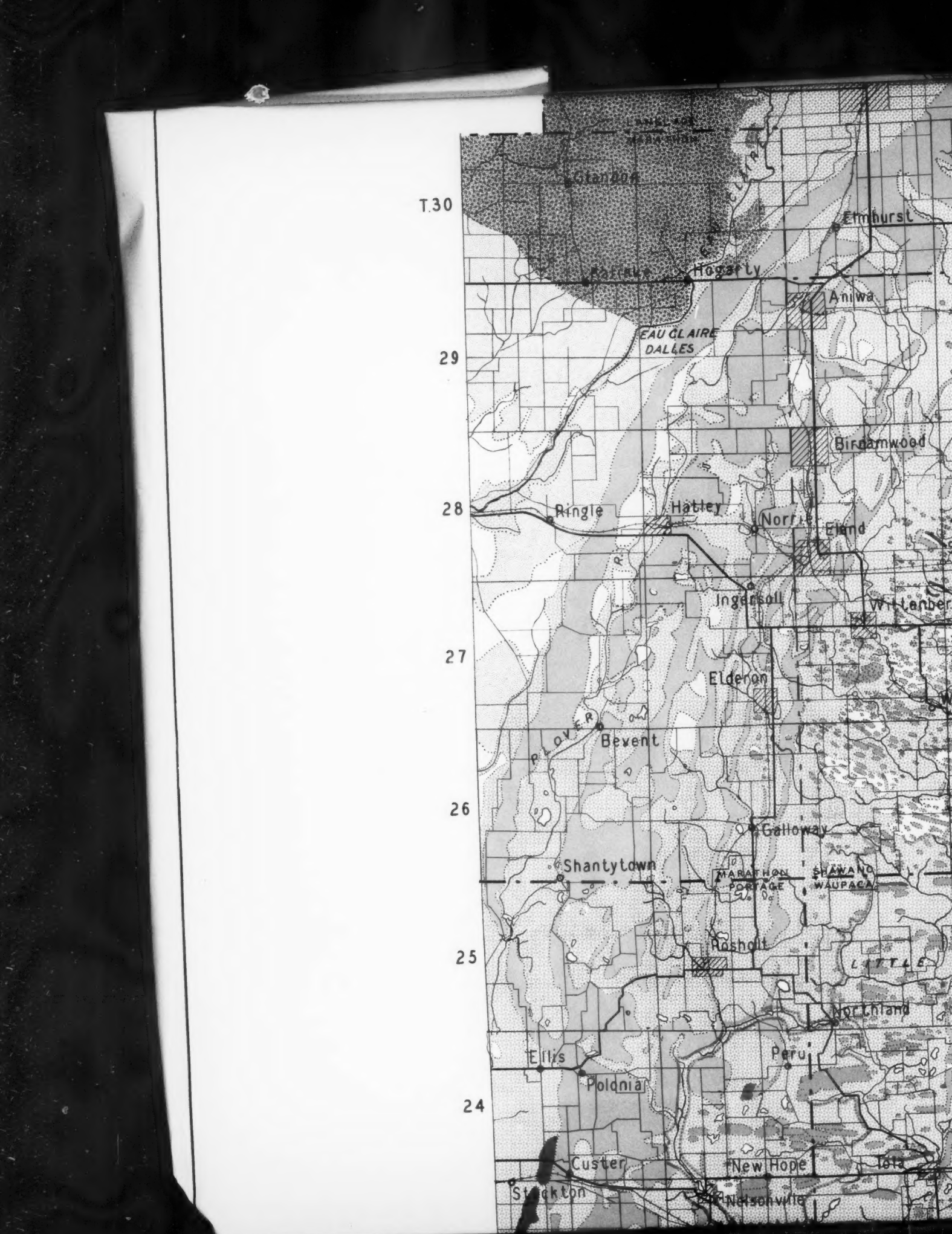
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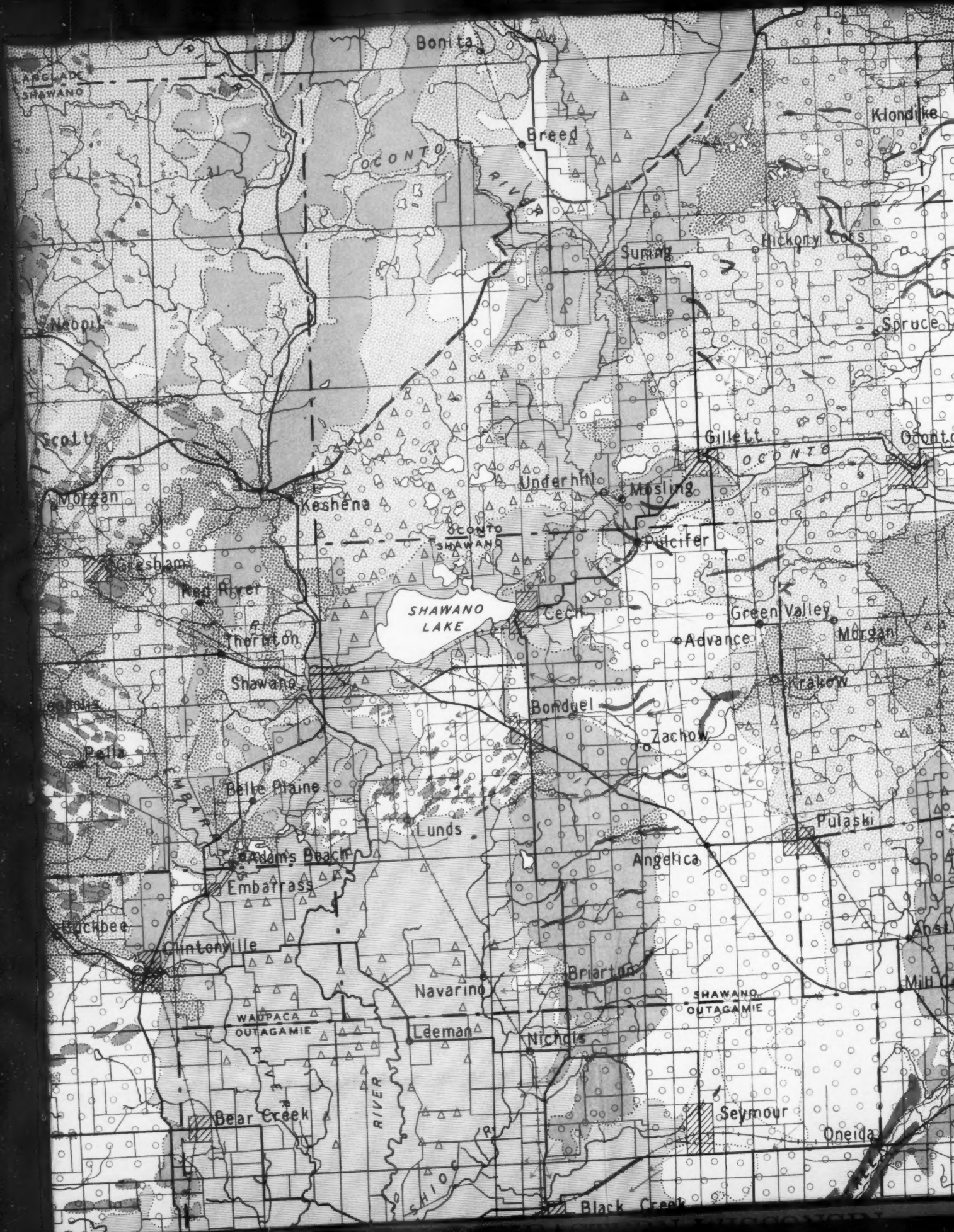
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Red River
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LAKE
Cecil
Green Valley
Morgan
Thornton
Advance
Shawano
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Pulaski
Adam's Beach
Angelica
Embarrass
Clintonville
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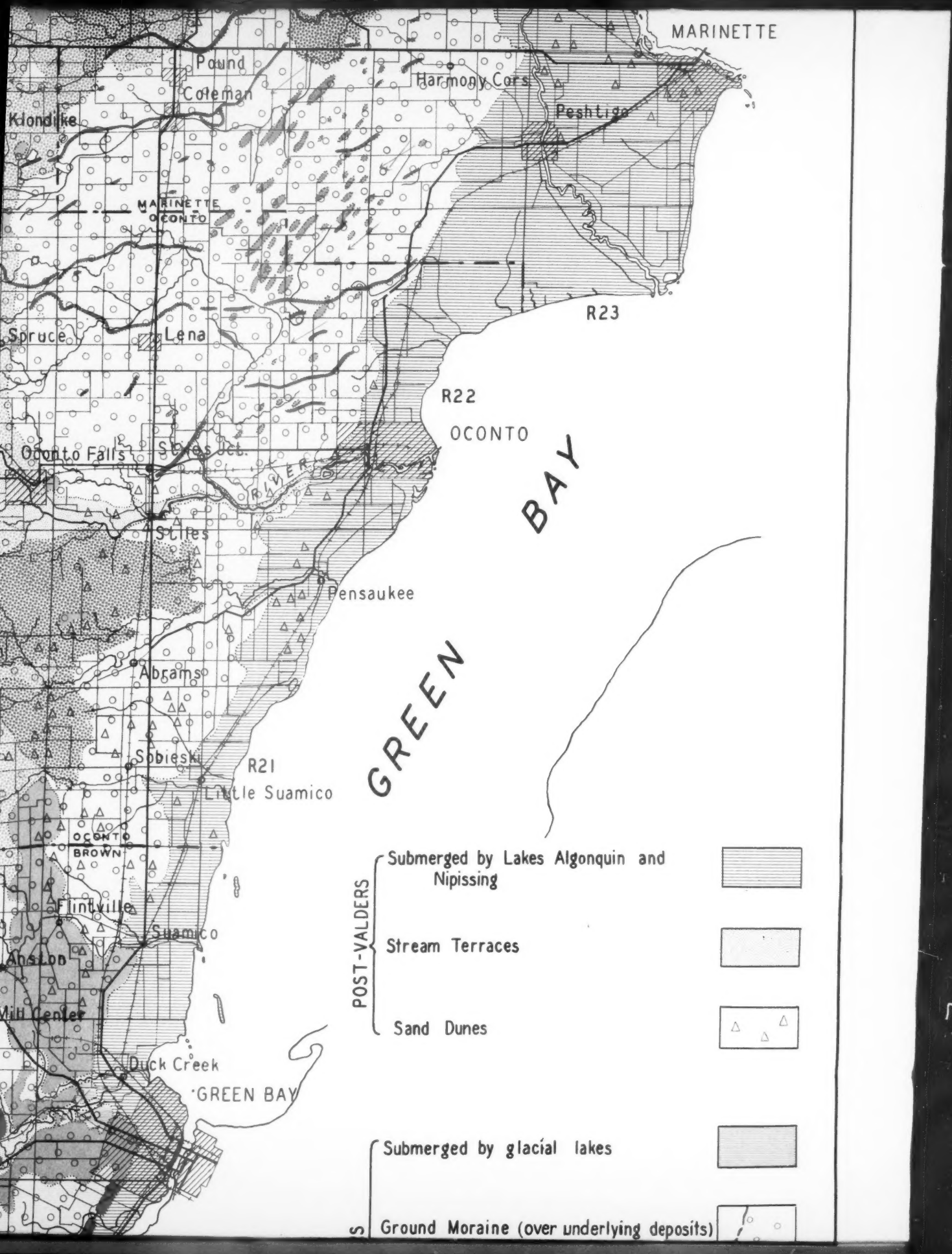
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Stiles Jct.
RIVER
Stiles
BAY
Pensaukee
GREEN
Abrams
Sobieski
R21
Little Suamico
OCONTO
BROWN
Flintville
Suamico
Ashlon
Mill Center
Duck Creek
GREEN BAY
POST-VALDERS
Submerged by Lakes Algonquin and Nipissing
Stream Terraces
Sand Dunes
Submerged by glacial lakes
Ground Moraine (over underlying deposits)

Custer
Stockton
New Hope
Tell
Annaliswhe
Amherst
Scandinavia
Fancher
Arnott
Amherst
23
Sherman
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Kiene
Rural
Blaine
T.21
Almond
PORTAGE
WAUSHARA
Plainfield
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Wild Rose
Saxeville
Pine
Hancock
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Red Gr
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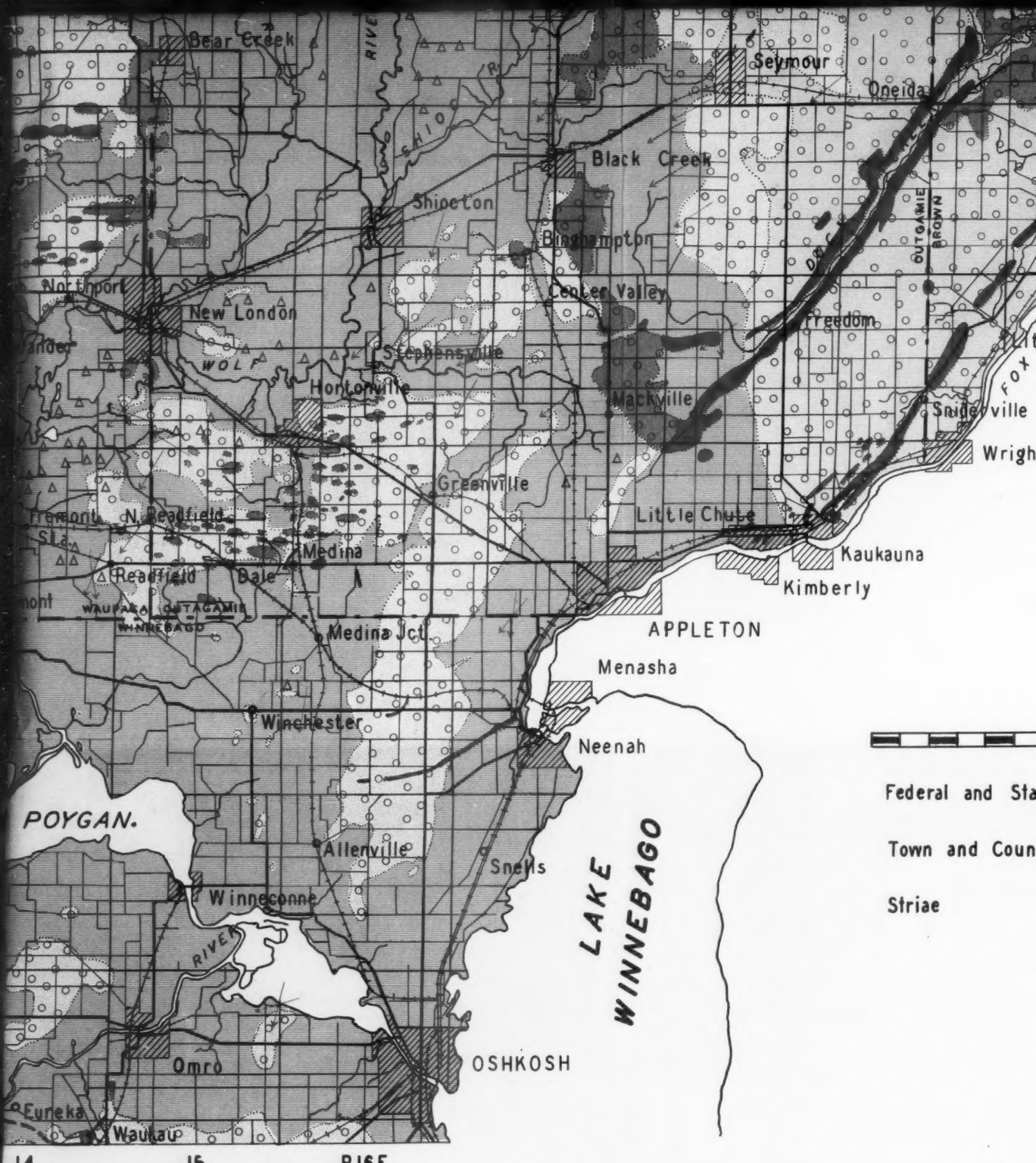
PLEISTOCENE GEOLOGY OF PART OF NORTHEASTER

F. T. THWAITES, 1926-1941



COLOGY OF PART OF NORTHEASTERN WISCONSIN

F. T. THWAITES, 1926-1941

GREEN BAY
De Pere
RIVER
lt. Rapids
FOX
ville
Wrightstown
10miles
nd State Highways
County Roads
WISCONSIN
VALDERS
Submerged by glacial lakes
Ground Moraine (over underlying deposits)
Outwash and Deltas
Eskers and Drumlins, in part Cary
CARY
Submerged by glacial lakes
Terminal Moraines and Deltas
Ground Moraine
Drumlins
Outwash and major terraces
Eskers
PRE-WIS. IOWAN?
PRE-CARY
Ground Moraine
ILLINOIAN?
Terminal Moraine
Thin or doubtful Drift - Colby border